# Bulletin

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## PAPERS IN PRESS FOR FORTHCOMING ISSUES

GEOLOGY OF THE HANOVER QUADRANGLE, NEW HAMPSHIRE AND VERMONT. By John B. Lyons

A geological map of the Hanover quadrangle, in color, accompanies this paper. Separate copies of the map, unfolded and in tubes, will be available. Price 25 cents. Remittance should accompany order. Maps will be available by December 15.

TRACE ELEMENTS IN PELITIC ROCKS. By Denis M. Shaw

Part I. Variations during Metamorphism

Part II. Geochemical Relations

MARINE GEOLOGY OF THE NORTHWEST PACIFIC: DESCRIPTION OF JAPANESE BATHYMETRIC CHART 6901. By Robert S. Dietz

Separate copies of the bathymetric chart, unfolded and in tubes, will be available. Price 50 cents. Remittance should accompany order. Maps will be available by December 15.

PERIDOTITE OF MANPUR, SINGHBHUM DISTRICT, BIHAR, INDIA, AND ORIGIN OF ASSOCIATED ASBESTOS DEPOSITS. By S. C. Chatterjee

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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# LOWER AND LOWER MIDDLE CAMBRIAN FAUNAL SUCCESSION IN NORTHWESTERN VERMONT

By Alan B. Shaw

## Abstract

The Gilman quartzite, Dunham dolomite, Monkton quartzite, and lower Parker slate all contain Lower Cambrian faunas, of which the trilobites *Olenellus, Bonnia, Pagetides, Kootenia marcoui,* and various ptychopariids predominate; among the brachiopods only *Kutorgina* and *Nisusia festinata* are common. The fauna of the Monkton is the correlative of the lower Parker fauna.

The upper Parker contains a lower Middle Cambrian (*Albertella* zone) fauna with *Zacanthoides, Sypacephalus, Kootenia, Chancia,* and other genera; all species are new except *Mexicella stator.*

The reinstatement of the name Georgian series to replace Waucoban series is urged.

Detailed descriptions and faunal lists of all fossil localities in the Lower and lower Middle Cambrian of northwestern Vermont are given.

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## Introduction

This paper and two others in preparation summarize present knowledge of the Early Cambrian and early Middle Cambrian faunas of northwestern Vermont. Although based upon original field work, this paper includes a thorough re-examination and collation of the older paleontologic studies. The writer offers two contributions: (1) the stratigraphic ranges of the species, and (2) the recognition of an early Middle Cambrian fauna at the top of the Parker slate, a formation heretofore believed to be wholly Lower Cambrian.

## Acknowledgments

W. A. Bell supplied information on several of Elkanah Billings' types, which are held by the Geological Survey of Canada. Otto Haas arranged for the loan of types from the American Museum of Natural History and supplied information on other types. G. A. Cooper provided a photograph of the cast of the holotype of *Protypus hitchcocki,* which is in the U. S. National Museum. R. C. Ramsdell, F. B. Van Houten, S. W. Lowman, J. L. Rosenholtz, and Winifred Goldring all aided my unsuccessful search for some of Ebenezer Emmons'

types. The late P. E. Raymond loaned his collections from the Parker slate. Christina Lochman Balk criticized the partly completed manuscript, and W. M. Cady thoroughly revised the finished paper.

Franco Rasetti assisted in the collection of part of the new Middle Cambrian fauna listed below and discussed certain nomenclatural problems. J. M. Boucher assisted in quarrying much of the Middle Cambrian fauna from the top of the Parker slate on his land.

Miss Diane Crawford typed drafts of the manuscript, and Miss Ann Kelsay typed the final manuscript under a grant-in-aid from the University of Wyoming Graduate Research Council.

B. F. Howell, A. R. Palmer, G. B. Maxey, S. W. Stone, and V. H. Booth contributed by discussion, but none must be charged with responsibility for the conclusions expressed herein.

## Stratigraphic Summary

Detailed discussions of the Lower Cambrian stratigraphy of west-central Vermont (Cady, 1945) and of the Oak Hill succession (Booth, 1950) have recently appeared, and Clark (1934; 1936) has summarized the same sequence in southern Quebec. These publications, supplemented by the writer's unpublished survey of the region around St. Albans, have clarified the Lower Cambrian succession, and especially the fossiliferous upper part including the Gilman quartzite, Dunham dolomite, Monkton quartzite, and Parker slate. Figure 4 shows the stratigraphic succession as it is now understood.

Gilman quartzite, the oldest fossiliferous formation, is conformable beneath Dunham dolomite, which is truncated by an erosion surface that in places cuts out as much as 2000 feet of the Dunham. On this erosion surface lies the Parker slate, a unit ranging from at least 1000 feet to as little as 25 feet thick where it has been cut by Middle Cambrian erosion.

The Monkton quartzite does not extend northward as far as St. Albans (Cady, 1945, map), but its fauna, described by Kindle and Tasch (1948), is like that of the lower Parker slate and, thus, warrants inclusion in this summary. The Winooski dolomite, which appears in the field to be conformable with the Monkton and which has lately been regarded as Lower Cambrian (Cady, 1945, p. 522), probably correlates with the Middle Cambrian Rugg Brook dolomite.

## Gilman Quartzite

### *General*

The most recent discussion of the Gilman is that by Booth (1950, p. 1148). The Gilman in the St. Albans area is a pale-green, tightly cemented, chloritic quartzite; weathered exposures are white or gray. It is not suited for the preservation of the small, delicate shells of most Early Cambrian species.

Clark (1936, p. 146) has noted the discovery of the heavy-shelled brachiopod *Kutorgina* near Scottsmore, Quebec, and, although the writer has found little more to add to the fauna, *Kutorgina* is sufficient indication of Early Cambrian age.

### *Fossil Localities*

LOCALITY SA-SE-3: Pale-green, slightly argillaceous quartzites about half-way up the middle of the east face of the quarry on Aldis Hill in the northern part of St. Albans have yielded three imperfect specimens of *Salterella*. In addition, these beds, which are probably about 100 feet below the top of the Gilman, contain comminuted limonitized trilobite fragments, none of which is generically identifiable but some of which seem to be derived from olenellids.

Figure 1 shows the location of fossil quarries described.

LOCALITY EF-WC-1: Dark-gray quartzites at the base of the transition zone between the Gilman and the Dunham at the 320-foot contour on the crest of the ridge 2.13 miles N. 79½°W. of Bench Mark 339 southwest of Sheldon Junction, in the Enosburg Falls quadrangle, yielded two specimens of *Hyolithes* sp. indet., one *Hyolithellus?*, and many indeterminable organic fragments.

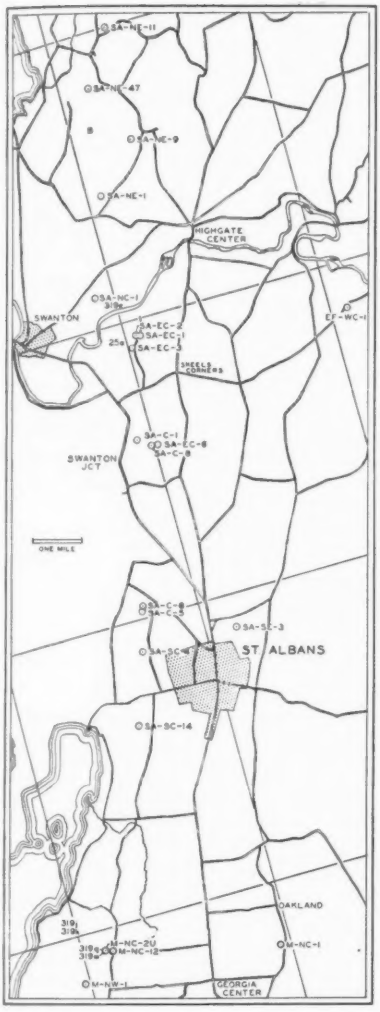


FIGURE 1.—SKETCH MAP OF NORTHWESTERN VERMONT SHOWING LOCATION OF FOSSIL COLLECTIONS

## DUNHAM DOLOMITE

### *General*

The Dunham dolomite in northwestern Vermont ranges from 25 to more than 2800 feet of gray and red dolomite, red arkosic sandstone, white quartzite, black shale, and blue-white limestone complexly interbedded. The two principal facies are indicated in Figure 2. The southern or carbonate facies is chiefly gray dolomite and magnesian limestone, but it contains some red, mottled dolomites that are quarried for ornamental stone. The northern or

FIGURE 2.—SCHEMATIC DIAGRAM OF FACIES RELATIONSHIPS IN DUNHAM DOLOMITE

sandy facies comprises mainly gray, highly arenaceous dolomite in the upper part and red sandstone in the lower. The red color is caused by finely divided arkosic fragments. Near the International Boundary the sandy facies is dominated by white, thick-bedded quartzite.

The thickness of the Dunham in the Rosenberg succession is extremely variable because the top was eroded during pre-Parker time and because the base is everywhere cut off by the Champlain thrust (Fig. 4). The base of the Dunham is exposed in the Oak Hill succession (Fig. 4), but there most of the upper part was eroded in pre-Parker time.

### *Fossil Localities*

The fossils from the Dunham in northwestern Vermont have been found largely in the St. Albans quadrangle, and even there they are scarce. Specimens were described first by Billings (1861, p. 3, 10, 11–13, 15) and have since been discussed and figured by Walcott (1886; 1891a), Resser (1937; 1938), and others.

Collections described here are located stratigraphically by distances downward from the top of the formation, since that level is less irregular, in spite of pre-Parker erosion, than is the lower limit, which is the plane of the Champlain thrust. Inasmuch as the relief on the upper Dunham surface is at least 50 feet, measurements show relative position only if they differ by more than that amount. In addition, because of the heavy and ubiquitous Pleistocene cover in this area the estimates of

stratigraphic position were derived trigonometrically and may be inaccurate by as much as 10–20 per cent.

Six localities described previously and four new localities are shown on Figure 1.

LOCALITY A: This locality was described by Billings (1861, p. 15) as being

> "in thin bedded, flaggy sandstone by the side of the road leading from Moore's Corners in St. Armand to Saxe's Mill in Highgate, Vermont, about one mile south of the Province line."

This outcrop was not rediscovered, but thin-bedded, flaggy sandstones are exposed only in the 20- to 25-foot sequence at the base of the section, immediately east of the road which Billings described and just north of the site of Saxe's Mill (marked by a plaque). If these sandstones are the site of Billings fossil quarry, then Locality A must lie about 550–600 feet below the top of the Dunham. The species reported here are *Perimetopus arenosus* (type locality) and *Planolites congregatus* (type locality).

Resser (1937, p. 51) garbled Billings' description of Locality A, saying that it lay "near the international boundary on the road from St. Armand to *Highgate Springs*" [italics mine]. Such a locality would be in the Ordovician rocks west of the Champlain thrust.

LOCALITY B: Another locality was described by Billings (1861) as "1 mile east of Highgate Springs." It was not recovered, but in the Museum of Comparative Zoology at Harvard College there is a collection, presumably from this locality, made by the Reverend Perry, who directed Billings to the outcrop originally. The matrix of the Perry collection is an arkosic red sandstone of a type present 600–1200 feet below the top of the Dunham; however, those sandstones most closely resembling the Perry collection are most common between 1000 and 1200 feet below the top of the formation, and it seems likely that Locality B lies somewhere in this interval. Species reported here is *Billingsaspis adamsii* (type locality).

USNM LOC. 26: The position of this locality is confused. The original description (Walcott, 1912b, p. 188) is: "Lower Cambrian: Sandstone northeast of the Corman farm buildings, east of Highgate Springs, Franklin County, Vt." Resser (1937, p. 49), in discussing *Ptychoparella adamsii* (= *Billingsaspis adamsii* of this report), states that it is "1 mile east of Highgate Springs," or the same as Locality B. Later, Resser (1938, p. 66) gave Locality 26 as "1½ miles southeast of Highgate Springs" in his comments on *Bonniella desiderata*. It is no longer possible to distinguish the correct from the incorrect citations in the literature, and USNM Locality 26 must be regarded either as impossible to locate geographically or as a composite locality. USNM Locality 26 may have been intended in part as a synonym of Locality B, but the original description does not make this certain, and the inclusion of the type locality of *B. desiderata* under this number adds to the confusion.

Inasmuch as the locality itself is under suspicion the fauna reported must likewise be regarded as possibly composite. The following species have been recorded:

*Antagmus typicalis* (type locality)
*A.? simplex* (type locality)
*Billingsaspis adamsii*
*Bonniella desiderata* (type locality?)
*Ptychoparella teucer*
*Nisusia festinata*
*Paterina labradorica swantonensis*
*Wimanella? orientalis*
*Scenella varians* (type locality?)
*Hyolithes* sp. indet.

USNM LOC. 319j: Walcott (1912b, p. 251) recorded a second faunule from "sandstone 50′ (15m.) above the base of the section west of Parker's quarry and about a mile (1.6 km) from the shore of Lake Champlain, in the township of Georgia, Franklin County, Vermont." This locality would be in the lowest beds above the Champlain thrust and must lie about 600 feet below the top of the Dunham. Fossils listed from this place include:

*Atops trilineata*
*Olenellus* sp. indet.
*Botsfordia caelata*
*Hyolithellus micans*
*Hyolithes americanus*
*H. communis*

It seems likely that the trilobite called by Walcott *Atops trilineata* would not now be identified with that species. Dr. B. F. Howell (Personal communication) likewise doubts the presence of *Atops* in northwestern Vermont. Similarly, *Botsfordia caelata* in the above list would probably now be differently named, for that species has become a wastebasket.

USNM LOC. 319k: Walcott (1912b, p. 251) recorded a collection from a "calcareous sandstone near the base of the section west of Parker's quarry, on cliff overlooking Lake Champlain, near Georgia, Franklin County, Vermont." This description might apply to any of several ledges ranging from 25 to nearly 200 feet stratigraphically above the Champlain thrust, or somewhere in the zone from 450 to 625 feet below the top of the Dunham, but the bed is probably at about the same level as USNM Locality 319j. Walcott listed *Obolella crassa* from this locality, but this species might be differently named in modern work.

LOCALITY SA-NE-47: Schuchert (1937, p. 1028) described a bed of red arkosic sandstone containing "*Ptychoparia*" *adamsi* and incorrectly assigned it to the Monkton formation. The writer revisited the locality, which is 0.93 mile S. 14½°E. of Bench Mark 99 just east of Rock River School, Highgate township, and discovered two cranidia which confirmed the presence of *Billingsaspis adamsii*. This bed lies about 515 feet below the top of the Dunham and a short distance north of Locality B.

LOCALITY M-NC-1: This locality, one of the four discovered in the present survey, is in the north-central ninth of the Milton quadrangle, 1380 yards, S. 9°W. of the three-corners at Oakland. The outcrops are along the west side of an elongate north-south ridge of dolomite. The fossils were collected from prominent 2-inch beds of argillaceous dolomite, now largely weathered to ochre. The relative position of this zone and those to the west and northwest is difficult to determine because post-Dunham erosion cut far deeper in this region than to the west. The writer estimates the fossils were collected 1000–1500 feet below the top of the Dunham, as it is exposed west of Georgia Center. The fauna at M-NC-1 includes: *Olenellus* sp. indet. (1 cranidium) and Olenellidae indet. (9 fragments).

LOCALITY SA-NE-11: Blue-gray coprolitic (?) limestone crops out in the bed of the small stream draining Proper Pond at 45°N. Lat. This bed is at the top of the Dunham and contains abundant *Hyolithes americanus*, but no trilobites or brachiopods were found.

LOCALITY SA-NE-1: A heavy-bedded, brown, quartzitic sandstone at the base of a west-facing ledge 1.78 miles N. 63°W. of Bench Mark 308 in Highgate Center, Highgate township, approximately 585 feet below the top of the Dunham, has yielded one valve of *Nisusia festinata*.

LOCALITY SA-NC-1: This locality is an old limestone quarry, badly overgrown by brush, 1.11 miles N. 60°½E. of the railroad crossing at elevation 155 at Swanton. This quarry exhibits one of the rare occurrences of pure blue-white limestone in the Dunham. The outcrop is surrounded by Quaternary deposits so that no contacts can be observed, but, by projecting the strike into better-exposed sections north or south, an estimate of some 500 feet below the top of the Dunham is obtained. The fauna obtained was sparse, but prospects for satisfactory collections seem good, if sufficient time were available for careful quarrying. Fauna noted:

*Olenellus* sp. indet. (1 free cheek; thoracic segment)
Dorypygidae indet. (1 cranidium)
*Hyolithes* sp. indet. (5 tubes)

### *Paleoecologic Observations*

Small though these collections are, they give some data on the ecologic preferences of some of the genera. The olenellid trilobites seem to have been unaffected by the type of bottom sediment, for they are found in limestone, in leached residues from dolomite, in sandstone, and in greatest abundance in black slate. The olenellid remains, however, have been seriously affected by the bottom conditions, for only in black slate, clearly deposited in undisturbed waters, are the olenellid tests preserved intact; elsewhere, the tests are fragmentary, and identifiable specimens are all but nonexistent. This selective preservation early suggested that the olenellids were confined to black muds and were, therefore, mud feeders. The wide discovery of olenellid remains suggests that this was not the case but that *Olenellus* and its relatives were swimmers, unaffected by any particular bottom conditions.

Like the Olenellidae, the Dorypygidae, including *Kootenia* and *Bonnia*, seem to have ranged over many kinds of bottom, but unlike the olenellids they seem to have been affected in some way by the conditions affecting

sedimentation, for the specimens preserved in black slate are usually smaller than those from carbonate rocks.

Unlike the Olenellidae and the Dorypygidae, the members of the Ptychopariidae (including *Billingsaspis*, *Ptychoparella*, and others) are largely confined to noncalcareous clastic sediments. Specimens are only rarely found in carbonate rocks. Individual species show marked confinement to a particular type of clastic lithology; for example, *Billingsaspis adamsii* is abundant in red arkosic sandstone, but specimens are rare from slate. The situation is reversed for *Ptychoparella teucer*. How much of this apparent preference for sandstone over shale, or *vice versa*, is due to ecologic effect and how much is due to replacement of one species by another in time cannot now be determined with certainty. Unfortunately no data on the abundance of species were recorded for the older collections.

## Parker Slate

### *General*

The Parker slate, named by Keith (1932, p. 371) from exposures around Parker Cobble, 2¼ miles west-northwest of Georgia Center, comprises black shales with interbedded local dolomite lenses. The formation is slightly more than 1000 feet thick 2 miles southeast of Swanton, but it thins both northward and southward where it was eroded in mid-Middle and mid-Late Cambrian times.

In Figure 3, a generalized columnar section based largely on the section southeast of Swanton, the Parker is divided nearly in two by a dolomite zone about 600 feet above the base; this dolomite is present in discontinuous lenses, which locally reach a thickness of 50 feet but are commonly thinner. Below the middle dolomite lies an argillaceous succession, here termed the *lower Parker slate*, which contains many local dolomite lenses and dolomite bioherms. The slates above the *middle dolomite* contain no dolomite and will be referred to as the *upper Parker slate*. Limestone bioherms are present in both lower and upper Parker slates but are rare.

The principal rock type in the Parker is a gray or black micaceous slate. Locally, the slate is extremely fine-grained, as at the old Noah Parker quarry where many soft-bodied organisms were preserved. Resser and Howell (1938) have given the most recent illustrations of these rare fossils.

Most of the fossil tests have been replaced by iron oxide, but some are preserved as impressions. Very few specimens, except the scarce inarticulate brachiopods, retain any trace of their original shell matter.

The dolomites are generally gray on fresh surfaces but weather brown; almost all except the bioherms are arenaceous. No trace of fossils can be found in fresh dolomite, but where the rocks have weathered to a sandy ochre there are trilobite coquinas. Presumably, therefore, all the dolomites are fossiliferous, but their fossils are invisible.

Some limestones in the lower Parker slate are highly fossiliferous, and collections in the Museum of Comparative Zoölogy from the "Hall Farm" show a coquina of *Kutorgina cingulata* shells with few trilobites. No fossils were found in the middle dolomite.

### *Fossil Localities in Lower Parker Slate*

Fourteen fossil localities are known in the lower Parker slates, dolomites, and limestones (Figs. 1, 3).

LOCALITY M-NW-1: Dr. P. E. Raymond made a large collection of fossils from the lower 10 feet or so of the Parker 0.50 mile N. 64½°W. of the crossroads (elevation 266) 2 miles west of Georgia Center, Georgia township, Milton quadrangle. The fossiliferous beds are orange-red sandy ochre, the residue from thin arenaceous dolomite beds. The fauna contains:

*Austinvillia* n. sp. (48 cranidia; 3 pygidia)
*Bonnia capilo* (156 cranidia; 159 pygidia; 32 hypostomata)
*Kootenia marcoui* (2 pygidia)
*Olenellus* sp. indet. (16 fragments, possibly representing no more than one individual)
*Zacanthopsis* sp. indet. (2 cranidia)
1 undetermined gastropod
1 undetermined crustacean (?)

LOCALITY SA-SC-14: Sandy residues similar to those at Locality M-NW-1 are present at the base of the Parker on the Conner Farm, 0.60 mile S. 15°W. of Bench Mark 220, on the road between St. Albans and St. Albans Bay. The Parker is here reduced by pre-Rugg Brook

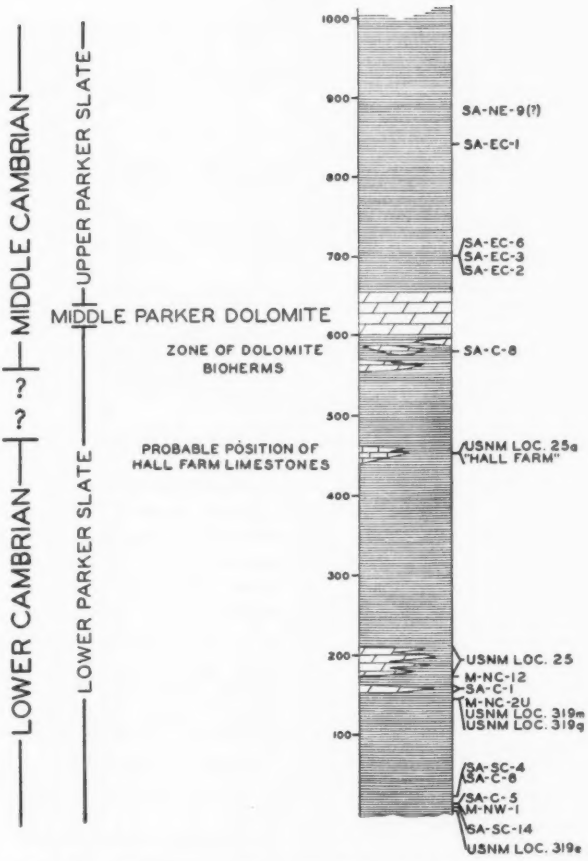


FIGURE 3.—GENERALIZED COLUMNAR SECTION OF PARKER SLATE

(mid-Middle Cambrian) erosion to a thickness of 25 feet. The fossiliferous dolomites lie in the lower 5 feet and contain:

*Kutorgina cingulata* (1 ventral; 1 dorsal (?) valve)
*Bonnia bubaris* (2 cranidia; 1 pygidial mold)

LOCALITY SA-SC-4: An outcrop of fossiliferous slate lies just east of the crest of a low north-south ridge, 0.15 mile N. 73°W. of the railroad crossing at elevation 343 at the northwest end of the St. Albans marshalling yards. The slate is about 20 feet above the base of the Parker and has yielded:

*Kootenia* sp. indet. (1 fragmental pygidium)
*Nisusia festinata* (1 valve found in float near this outcrop)

LOCALITY SA-C-5: Fossils are common in the matrix of a conglomerate at the base of the Parker on the eastern side of the Central Vermont Railroad tracks 0.62 mile S. 8°E. of Tuller School, St. Albans township. The matrix of the conglomerate is a gray micaceous slate, and the pebbles are Dunham dolomite. The top of the Dunham is channeled at least 8 feet. The fossils are 8–10 feet above the bottoms of the deepest channels or about on a level with the tops of the highest pinnacles of Dunham dolomite. The fauna contains:

*Bonnia capito?* (2 pygidia)
*Ptychoparella teucer* (2 cranidia)
*Nisusia festinata* (1 valve)
*N. transversa* (1 ventral valve)

*Nisusia?* (2 fragments)
cystid plate (?) (1 specimen)

LOCALITY SA-C-6: A pile of fine-grained black slate removed during widening of a railroad cut 0.51 mile S. 16°E. of Tuller School yielded one cranidium of *Billingsaspis adamsii*. This cut was new in 1946, and there seemed to be no contamination by rock from elsewhere. The cut is about 750 feet north of Locality SA-C-5 and about 20 feet above the base of the Parker.

USNM LOCALITY 319e: Walcott (1912b, p. 250) gave this number to Billings' old locality "1½ miles east of Swanton, in slates of the Potsdam group." No such outcrop was rediscovered, but it must have been close to the base of the Parker, probably along the north bank of the Missisquoi River, for this is the only place where slates might be found 1½ miles east of Swanton. We may tentatively regard USNM Locality 319e as about in the stratigraphic position of the four just described. The fauna reported by Walcott included:

*Kutorgina cingulata*
*Nisusia festinata* (type locality)
*Swantonia antiquata* (type locality)
*Ptychoparella kindlei* (type locality)

USNM LOCALITY 319g: This locality was reported by Walcott (1912b, p. 250) as the "shales at Parker's quarry." The only species reported were: *Wimanella? orientalis* (type locality) and *Olenellus thompsoni*, but, since Emmons' specimens almost certainly came from the Parker quarry, this is probably also the type locality of *Olenellus brachycephalus*.

The writer concludes from a study of the published descriptions and an examination of Parker Cobble itself that the old Noah Parker quarry, which is now destroyed, lay in the shale series above a zone of bioherms on the west slope of the Cobble and below the arenaceous dolomite that caps the Cobble. This shale is 120–170 feet above the base of the formation, and the new collections made in the area suggest that the Parker quarry was near the middle of the interval. USNM Locality 319g and the four numbered localities that follow are all at Parker's Cobble.

USNM LOCALITY 319m: This locality, described by Walcott (1912b, p. 251) as "shales of No. 6 of the section at Parker's quarry, near Georgia, Franklin County, Vermont," is almost certainly the same as USNM Locality 319g. The reported fauna is the typical Georgian fauna:

*Kutorgina cingulata*
*Nisusia festinata*
*N. transversa*
*Wimanella? orientalis*
*Planolites congregatus*
*P. virgatus*
*Anomalocaris? emmonsi* (type locality)
*Emmonsaspis cambrensis*
*Bathynotus holopygus* (type locality)
*Billingsaspis adamsii*
*Bonnia capito*
*Kootenia marcoui* (type locality)
*Olenellus thompsoni* (type locality)
*O. vermontanus* (type locality)
*Pagetides parkeri* (type locality)
*Protypus hitchcocki* (type locality)

This locality is sometimes confused with USNM Locality 25.

LOCALITY M-NC-2U: This locality is directly above the bioherms on the southwestern slope of Parker Cobble, 990 feet N. 88°W. of the road interesection east of the Cobble, approximately 140 feet above the base of the Parker. Its fauna suggests that it is an extension of the old Parker quarry horizon. The fauna includes:

*Bonnia capito* (2 cranidia; 2 pygidia)
*Pagetides parkeri* (1 pygidium)
*Periomma typicalis* (1 cranidium)
*Olenellus* sp. indet. (2 fragments)

LOCALITY M-NC-12: This is 550 feet S. 63°W. of the road intersection mentioned above. It is 170 feet above the base of the formation and contains only *Nisusia festinata* (4 dorsal valves; 3 ventral valves) and *Paterina labradorica swantonensis* (1 valve).

USNM LOCALITY 25: This is stratigraphically the highest of the five localities described by Walcott (1912b, p. 188). He referred to it as "sandstone just above Parker's quarry . . .", This certainly refers to the arenaceous dolomite or dolomitic sandstone that caps Parker Cobble and extends down the eastern slope. No new fossils have been collected from this unit, which lies 170–210 feet above the base of the Parker. The reported fauna is:

*Rustella edsoni* (type locality)
*Paterina bella*
*Nisusia festinata*
*N. transversa* (type locality according to Walcott, 1912b, p. 188, but original description suggests USNM Locality 319m)
*Tuzoia vermontensis* (type locality)
*Eocystites?* sp.

*Olenellus vermontanus* (erroneously listed as type locality by Walcott, 1912b, p. 188, and by Resser and Howell, 1938, p. 200)
*O. thompsoni* (erroneously listed as type locality by Walcott, 1912b, p. 188, and by Resser and Howell, 1938, p. 221)
*O. brachycephalus* (type locality for synonymous species *Paedeumias transitans*)
*Pagetides parkeri*
*Billingsaspis adamsii*
*Bonnia capito?*
*Bathynotus holopygus*
*Kootenia marcoui*
*Protocaris marshi* (type locality)

So much confusion has arisen concerning this locality and USNM Locality 319m that it is no longer possible to separate the faunas of the two. For example, Walcott's discussion of the Mesonacidae (1910, p. 339) places the type locality of *Olenellus thompsoni* "in argillaceous shales of Parker's quarry, near Georgia," which was stated to be Locality 25. Two years later Walcott (1912b, p. 251) listed Locality 319m as the type, while Locality 25 was stated to be above the quarry (Walcott, 1912b, p. 188). Resser and Howell (1938, p. 209) seem not to have distinguished the two localities; they list the faunas of both as from Locality 25 (for example, the type locality of *Anomalocaris? emmonsi* is stated to be Locality 25 by Resser and Howell (1938, p. 232) and Locality 319m by Walcott (1912b, p. 251). Resser (1937) also added to the confusion by shifting locality assignments without adequate knowledge of the area.

Therefore, for all practical purposes the listed faunas for Localities 25 and 319m must be regarded as inextricably mixed, but probably most of the species now regarded as coming from Locality 25 (sandstone) actually came from Locality 319m (slate). The confusion cannot be resolved by field investigation because the Noah Parker quarry has been entirely quarried away.

Walcott (1886, p. 15) estimated the thickness of his Unit 6, which included both the slates and the dolomite capping the Cobble, at 200 feet; this agrees even more closely than might have been expected with the estimate of 210 feet obtained during this study. Schuchert (1937, p. 1013), in discussing this section, states that Unit 6 was "nearer 300" feet thick, but this does not seem correct.

LOCALITY SA-C-1: The only other locality representing the Parker quarry horizon is 0.58 mile N. 56½°E. of the railroad crossing at Swanton Junction. This is the less well known but highly fossiliferous Kelly quarry described by Schuchert (1937, p. 1035; Loc. XXI on map, p. 1004) and is notable for the great number of specimens of *Dactyloidites asteroides* exposed on beds at the entrance to the quarry.

There are two fossiliferous rock types at the Kelly quarry. The lower one is the black slate zone in the quarry proper, which lies an estimated 150–160 feet above the base of the Parker. Above this is a zone of brown-weathering, gray, nonarenaceous dolomite, which yields many specimens of trilobites on weathered surfaces. The dolomite is locally thick, but the fossils are confined mainly to the lower 5 feet, 160–165 feet above the base of the Parker. The fauna of the Kelly quarry slates is:

*Kootenia marcoui* (1 hypostoma; 1 articulated individual)
*Bathynotus holopygus?* (1 pygidium)
*Olenellus brachycephalus* (6 cephala)
*O. vermontanus* (1 glabella)
*Olenellus* sp. indet. (21 fragments)
*Pagetides parkeri* (39 cranidia; 46 pygidia)
*Prozacanthoides* sp. indet. (1 cranidium)
*Ptychoparella teucer* (2 entire individuals; 36 cranidia; 1 free cheek)
*Dactyloidites asteroides* (abundant)
*D. edsoni* (type locality abundant)

The fauna of the Kelly quarry dolomite includes:

*Kootenia marcoui* (23 cranidia; 27 pygidia)
*Olenellus* sp. indet. (1 fragment)
*Pagetides parkeri* (1 cranidium; 2 pygidia)
*Ptychoparella teucer* (1 cranidium)

Schuchert (1937, p. 1035) estimated the thickness of the Parker in the Kelly quarry section at 180 feet, but he and Keith mistook a local dolomite within the Parker (Unit 2a of Schuchert's Fig. 9) for the Mallett (= upper Dunham) repeated by thrusting and, therefore, failed to understand that the Parker thickens northward and is about 700 feet thick at this latitude. This is but one of the many instances in which the earlier workers missed major changes in the stratigraphic sequence, because they did not trace the units laterally but relied on the traditional methods of surveying successive parallel stratigraphic sections and correlating between them.

Above the Kelly-Parker quarry zone is

about 250 feet of black slate with local intercalated dolomite lenses. Fragments of slate float containing organic remains suggest that this sequence is fossiliferous, but no collections have been made from rocks *in situ*.

USNM LOCALITY 25a: This lower Parker fossil locality was described by Walcott (1912b, p. 188) as "limestone on the Hall (Donaldson's) Farm, 2 miles (3.2 km) east of Swanton, Franklin Count . . .". Resser (1937, p. 52), in his discussion of *Ptychoparella walcotti*, described the rock as a gray limestone and stated that "besides trilobites, the rock contains many thick-shelled *Kutorgina*." Specimens in the Museum of Comparative Zoölogy, whose labels show that they were collected on the Hall Farm by Reverend Perry, seem to have come from this spot. The rock is gray to dark-gray limestone showing little bedding; it is a coquina of shells of *Kutorgina cingulata* with few trilobites. This collection was not prepared, but the following count of surface specimens was made:

*Kutorgina cingulata* (7 dorsals; 8 ventrals)
*Billingsaspis walcotti* (11 cranidia)
*Ptychoparella teucer* (1 cranidium)

The Hall Farm quarry was not relocated during the present survey, but it certainly is not "two miles east of Swanton," as Walcott said, for such a locality would be in the Missisquoi River. The rock type most closely resembles the few limestone bioherms seen in the Parker, and these bioherms are confined to a small area bout 2 miles southeast of Swanton, in the rectangle bounded by 44°54′15″ and 44°54′50″ N. Lat., and by 73°5′00″ and 73°5′30″ W. Long. Contrary to statements made by Resser (1937, p. 52), the Hall Farm limestone does not resemble any part of the Dunham dolomite; therefore, the most likely estimate for the Hall Farm locality is that it is about 2 miles southeast of Swanton in the zone of lower Parker limestone bioherms, probably about 450 feet stratigraphically above the base of the Parker. If the ledge should ultimately be relocated, its stratigraphic position will almost certainly not be much more than 500 feet above the base of the formation because that is the upper limit of known limestone in the lower Parker. Possibly the position may be lower than that cited, but not over a couple of hundred feet because limestone is absent in the lower parts of the Parker, too.

USNM Locality 25a is not where Schuchert (1937, p. 1034) said it was. Schuchert's Locality XVIII (described below as Loc. SA-EC-1), which he regarded as the same as Locality 25a, is in the upper Parker slate where there is but a single limestone bioherm, wholly unfossiliferous and of different lithology from the old Hall Farm bed. Dr. P. E. Raymond first called this confusion to the writer's attention, and field work and collections have corroborated his opinion. Moreover, in a discussion in June 1954, Mr. Donaldson, the present owner, stated that Locality XVIII is not on the original Hall Farm at all. Finally, the Hall Farm fauna is clearly Lower Cambrian, whereas the fauna at Locality SA-EC-1 is Middle Cambrian.

The fauna reported from USNM Locality 25a includes:

*Paterina labradorica swantonensis* (type locality)
*Lingulella granvillensis*
*Nisusia festinata*
*N. transversa*
*Wimanella? orientalis*
*Hyolithes americanus*
*Salterella pulchella*
*Olenellus thompsoni*
*Billingsaspis adamsii*
*Bonnia senecta* (?)

LOCALITY SA-C-8: The highest collection found in the lower Parker was taken at the foot of the spur 0.71 mile N. 85°E. of the railroad grade crossing at Swanton Junction, from black slate estimated to be about 580 feet above the base of the Parker. This collection is characteristic of the upper Parker fauna in that it lacks *Olenellus*. Although the collection is small, the absence of even a single recognizable fragment of an olenellid, such as are abundant in the lower rocks even where no identifiable complete specimens have been obtained, seems to be good evidence that *Olenellus* was not present at this level. Also, the relative abundance of *Nisusia* suggests that we have here the upper fauna although the dominant species, *N. festinata*, is that of the lower Parker. The collections from Locality SA-C-8 and all higher localities are regarded as Middle

Cambrian. The fauna of Locality SA-C-8 includes:

*Nisusia festinata* (3 dorsal valves; 3 ventral valves)
*Nisusia*, n. sp. (1 ventral valve)
*Ptychoparella* sp. indet. (2 cranidia)
*Hyolithes americanus* (3 specimens)

### *Fossil Localities in Upper Parker Slate*

Five collections have been made from the upper Parker slate—three near the base and two near the top. In the interests of precision, the position of the first four is referred to the top of the middle dolomite rather than to the base of the formation. The stratigraphic position of the last, Locality SA-NE-9, cannot now be determined.

LOCALITY SA-EC-3: This exposure of gray-black micaceous slate lies on a low ridge 1.02 miles N. 46½°W. of Skeels Corners and is about 50 feet above the middle Parker dolomite. The fossils collected include:

*Nisusia festinata* (1 dorsal valve; 1 ventral valve)
*"Finkelnburgia"* sp. indet. (1 ventral valve)

LOCALITY SA-EC-2: This is a northward extension of the zone exposed at Locality SA-EC-3, and it is 1.11 miles N. 32½°W. of Skeels Corners. The species found were:

*Nisusia* aff. *N. amii* (4 valves)
*Kootenia* sp. A (1 cranidium; 1 pygidium; 1 free cheek)

LOCALITY SA-EC-6: A flat glaciated exposure in a large field 0.87 mile N. 82°E. of the railroad grade crossing at Swanton Junction shows slate also approximately 50 feet above the top of the middle dolomite. The fauna is:

*Nisusia festinata* (2 ventral valves)
1 cystid columnal

LOCALITY SA-EC-1: Schuchert (1937, p. 1034, and map, p. 1004) described this locality as his Locality XVIII, and it has become known as the *Donaldson quarry*. It is 1.03 miles N. 29½°W. of Skeels Corners and is stratigraphically the highest and one of the most fossiliferous exposures of the Parker slate. It is a single ledge, 10–15 feet long, of black, limonitic, micaceous slate. The outcrop can be recognized by the piles of loose rock left by several generations of collectors. The bed is 190 feet above the top of the middle dolomite and bears a Middle Cambrian fauna:

*Nisusia festinata* (2 ventral valves, 2 dorsal, and 1 unrecognizable)
*N.* aff. *N. amii* (10 ventral valves, 10 dorsal, and 1 unrecognizable)
*Nisusia* sp. indet. (2 valves)
*"Finkelnburgia"* sp. indet. (1 dorsal valve)
undetermined orthoid (2 dorsal valves)
*Hyolithes americanus?* (abundant)
*Helcionella* aff. *H. paupera* (3 shells)
*Kootenia* sp. A (3 cranidia, 8 pygidia)
*Mexicella stator* (2 cranidia)
*Syspacephalus* sp. A (3 cranidia)
1 undetermined ptychopariid trilobite cranidium
7 sponge spicules
4 cystid plates

After the bulk of the specimens from this locality had been prepared a small quantity of unworked slate was brought into the laboratory and prepared as carefully as possible to serve as a check on previous collections and to give a less biased sample of the species present. The distribution of this sample probably shows better the relative proportions of the species. Note the dominance of nontrilobitan elements:

*Nisusia festinata* (1 ventral valve; 1 unrecognizable valve)
*N.* aff. *N. amii* (6 dorsal valves, 5 ventral valves)
*"Finkelnburgia"* sp. indet. (1 dorsal valve)
*Hyolithes americanus?* (9 specimens)
*Helcionella* aff. *H. paupera* (2 shells)
*Kootenia* sp. A (2 cranidia, 2 pygidia)

These figures are included in the first list given above.

LOCALITY SA-NE-9: A single outcrop of gray and blue-gray limestone conglomerate was found 1.89 miles N. 22°W. of Bench Mark 308 in Highgate Center. This outcrop yields an early Middle Cambrian fauna, as was first recognized by Dr. Franco Rasetti, who kindly assisted in collecting. The conglomerate was at first considered as an independent formation, but when the same fauna was recognized in the upper Parker slate at Locality SA-EC-1 it was assigned to the Parker. It is impossible to refer Locality SA-NE-9 to the column given in Figure 3 because the outcrop is isolated, but its faunal content indicates that it is essentially the equal of Locality SA-EC-1.

The fauna of this locality includes:

*Athabaskia* sp. indet. (1 pygidium)
*Chancia* n. sp. (28 cranidia, 4 pygidia)
*Corynexochides* sp. indet. (1 cranidium)

*Kochina* sp. indet. (4 cranidia)
*Kootenia* sp. A (49 cranidia, 53 pygidia, 27 hypostomata, 2 free cheeks, 1 protaspis tentatively assigned to this species)
*Kormagnostus* sp. indet. (2 pygidia)
*Mexicella stator* (21 cranidia)
*Oriella* sp. indet. (4 cranidia, 7 pygidia)
*Peronopsis* sp. indet. (3 cephala, 2 pygidia)
*Syspacephalus* sp. A (103 cranidia)
*Zacanthoides* n. sp. (19 cranidia, 14 pygidia)
1 undetermined asaphiscid

This list includes all forms collected save the inarticulate brachiopods, which are numerous. Articulate brachiopods, the most conspicuous elements at Locality SA-EC-1, are absent. The writer assumes this is an ecologic effect.

Listing of the fossil localities in the Parker according to their stratigraphic position brings out a hitherto unrecognized faunal difference between the lower and upper parts of the formation. The most important change is the disappearance of olenellids at and above the 580-foot level, and the introduction of a fauna dominated in slate by *Nisusia* and other brachiopods and in limestone by *Kootenia* and other trilobites. The presence of *Mexicella*, *Chancia*, *Corynexochides*, *Oriella*, and of six-spined *Kootenia* tails indicates that the fauna is Middle Cambrian.

This faunal change has been overlooked previously because the genera in the upper fauna are essentially those of the lower. Furthermore, Schuchert's incorrect faunal list for the Middle Cambrian Donaldson quarry (Loc. SA-EC-1), which he mistook for the Lower Cambrian Hall Farm quarry (USNM Loc. 25a), masked the change by including *Olenellus* (Schuchert, 1937, p. 1034). Kindle (1942, p. 635) hinted at the presence of a post-olenellid fauna in the Parker but did not elaborate.

The Middle Cambrian part of the Parker slate is not a separate formation because it is not mappable. If emphasis is placed on the presence of *Mexicella stator*, the upper Parker fauna may be provisionally correlated with the *Albertella* zone as outlined by Lochman (1948, p. 451–452) and Rasetti (1951, p. 93–96).

## Monkton Quartzite

Fossils have been discovered only recently in the Monkton quartzite (Kindle and Tasch, 1948). Cady (1945, p. 531) concluded that "the Monkton lies at about the horizon of the Parker slate," and the work of Kindle and Tasch confirmed this.

The collection described by Kindle and Tasch was apparently made about 50–100 feet above the base of the Monkton (*compare* their Fig. 1 with Cady's areal map, 1945, Pl. 10), which thins in this area and disappears 3–4 miles to the north.

With the more precise faunal zoning of the Parker, the Monkton fauna can now be correlated with the Kelly-Parker horizon of the lower Parker fauna. This persistence of the lower Monkton beds, accompanied by the northward thinning of the formation, strongly suggests that the Monkton disappears by the removal, toward the north, of the top of the formation in a manner identical with that in which the Parker has been eroded southward. This situation also suggests that the Winooski dolomite, which overlies the Monkton and lies above the post-*Albertella* erosion surface on the top of the Parker, is the correlative of the Rugg Brook dolomite, which also lies above the post-Parker erosion surface. The similarity of the two dolomites also supports this correlation.

The Monkton fauna, with generic and specific assignments changed to agree with the nomenclature used in this paper, contains the following species:

*Olenellus brachycephalus*
*O. hermani*
*Billingsaspis adamsii*
*Bonnia capito*
*Kutorgina cingulata*
*Nisusia* sp. indet.

Other incomplete fragments of less certain identification are not listed.

## Faunal Zones

### *Lower Cambrian*

The Cambrian Subcommittee (Howell *et al.*, 1944) divided the Lower Cambrian into four faunal zones, reading upward, the *Obolella*, *Bonnia*, *Olenellus*, and *Syspacephalus* zones. No fossiliferous rocks which might be assigned to the lowest zone are known in northwestern Vermont, but units assigned to the upper three may serve as a basis for their reconsideration.

The refinement of the data on stratigraphic distribution of the Lower Cambrian genera points out clearly that *Bonnia* has its principal development in the lower Parker, in the same beds that contain abundant *Olenellus*; therefore, the writer suggests that the *Bonnia* and *Olenel-*

*lus* zones are not successive but equivalent and that the *Bonnia* zone should be abandoned. Rasetti (1951, p. 82) has reached this same conclusion from other evidence.

*Bonnia* zones (Fig. 4) of the Subcommittee and to the *Antagmus-Onchocephalus* zone of Lochman. It is not equivalent to the *Syspacephalus* zone of the Subcommittee. Rasetti (1951, p. 84ff.)

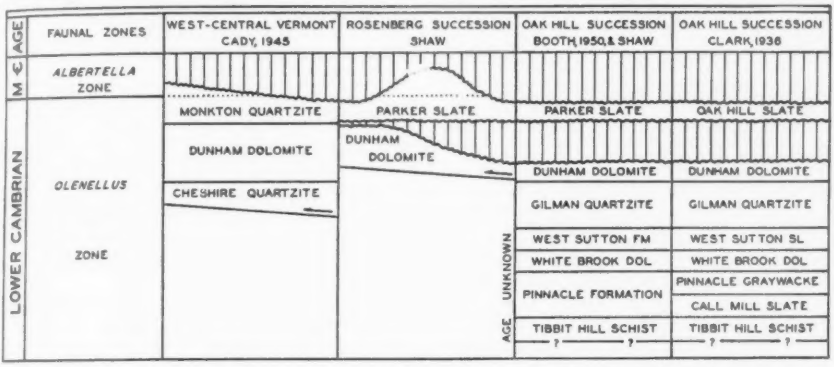


FIGURE 4.—CORRELATION OF LOWER CAMBRIAN FORMATIONS OF WESTERN AND NORTHWESTERN VERMONT AND SOUTHERN QUEBEC

Resser (1938, p. 20) and Resser and Howell (1938, p. 199) called the topmost Lower Cambrian zone the *Kochiella* zone, but the Cambrian Subcommittee chose *Syspacephalus* as the typical genus. Later, Lochman (1947, p. 62) suggested the name *Antagmus-Onchocephalus* zone be substituted for *Syspacephalus* zone because *Syspacephalus* was not a widespread genus.

The *Syspacephalus* zone was originally defined as a zone of late Early Cambrian faunas without *Olenellus*, but it appears from later articles that this is not what Lochman intended to describe when she proposed the name *Antagmus-Onchocephalus*. Instead, she was naming a zone in which olenellids are accompanied by other trilobites, as opposed to a restricted *Olenellus* zone which contained only *Olenellus* or its relatives, without other trilobites.

If Lochman's intended procedure is followed (*i.e.*, to set apart a segment of the *Olenellus* zone, as defined by the Cambrian Subcommittee, as the *Antagmus-Onchocephalus* zone), it means that the *Olenellus* zone (restricted) does not exist in northwestern Vermont. This places the type locality of *Olenellus thompsoni* itself out of the *Olenellus* zone. Therefore, the writer has only recognized a single *Olenellus* zone in northwestern Vermont, and that zone is equivalent to the combined *Olenellus* and has dealt thoroughly with the problem of post-*Olenellus* rocks and has concluded that all should be assigned to the Middle Cambrian, a procedure in which the writer fully concurs.

### *Early Middle Cambrian*

Rasetti (1951) has recognized two faunal zones beneath the *Albertella* zone, which Lochman (1948) regarded as the oldest Middle Cambrian. These are the *Wenckchemnia-Stephenaspis* and *Plagiura-Kochaspis* zones. Neither zone is recognizable in Vermont, but this does not seriously reflect on their validity, for that part of the Parker slate in which such faunas might be sought has not so far yielded many specimens. The 400-foot interval between the clearly Lower Cambrian Hall Farm collection (USNM Loc. 25a) and the clearly Middle Cambrian Locality SA-EC-1 has produced only 25 specimens from two levels; this is obviously insufficient evidence either to establish or refute the presence of the *Wenckchemnia-Stephenaspis* and *Plagiura-Kochaspis* zones.

The fauna at Locality SA-NE-9 contains 11 genera of trilobites whose cumulative range extends from the *Olenellus* zone possibly as far up as the *Thompsonaspis* zone (late Middle Cambrian). The recorded ranges of the genera cannot be reconciled with the presence of all

the genera at one locality, but the dominant forms are of early Middle Cambrian age; the writer suggests provisional correlation with the *Albertella* zone. In this dating emphasis has been placed on *Mexicella*.

## Series Terminology for Lower Cambrian

Walcott (1891a) used the name *Olenellus* zone essentially in the sense of a series term after he had established the correct relative positions of the *Olenellus* and *Paradoxides* faunas. Later, he (1891b, p. 360) applied the formal name Georgian series to the Lower Cambrian, referring to a standard section described by him (1886, p. 15–18) west of Georgia Center, Vermont. The name Georgian was replaced, about 20 years later, by Waucoban series (Walcott, 1912a, p. 305–305) because Georgian was at that time also used for a formation, the Georgia slate.

The writer visited Walcott's Waucoba section with Dr. G. B. Maxey and Mr. J. C. Bollenbacher, in August, 1952, a trip made possible by a Grant-in-aid from the Yellowstone-Bighorn Research Association. This examination, although rapid, suggested that the Waucoba section is not, in our present state of knowledge, a satisfactory standard and that we should return to the use of the term Georgian series for the following reasons: (1) The Georgia section, as opposed to that at Waucoba Springs, is now well known, and its contained fossils are properly located stratigraphically. (2) The fauna of the Georgia section, and especially of the Parker slate, is varied and abundant. On the other hand, fossils are not common and do not seem to be so varied in most of the Waucoba section. (3) The name Georgian is no longer in use for any other stratigraphic unit.

On the other hand the Waucoba section is two or three times longer than that in Vermont, and the term Waucoban series has been used for 40 years. However, the length of the Waucoba section is of little practical advantage when so little is known of either its detailed stratigraphy or its faunas.

The important question is whether or not the standard section has a fauna well enough known and large enough to permit interregional correlations. This is true of the Georgian fauna but not of the Waucoban fauna.

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Univ. of Wyoming, Laramie, Wyo.

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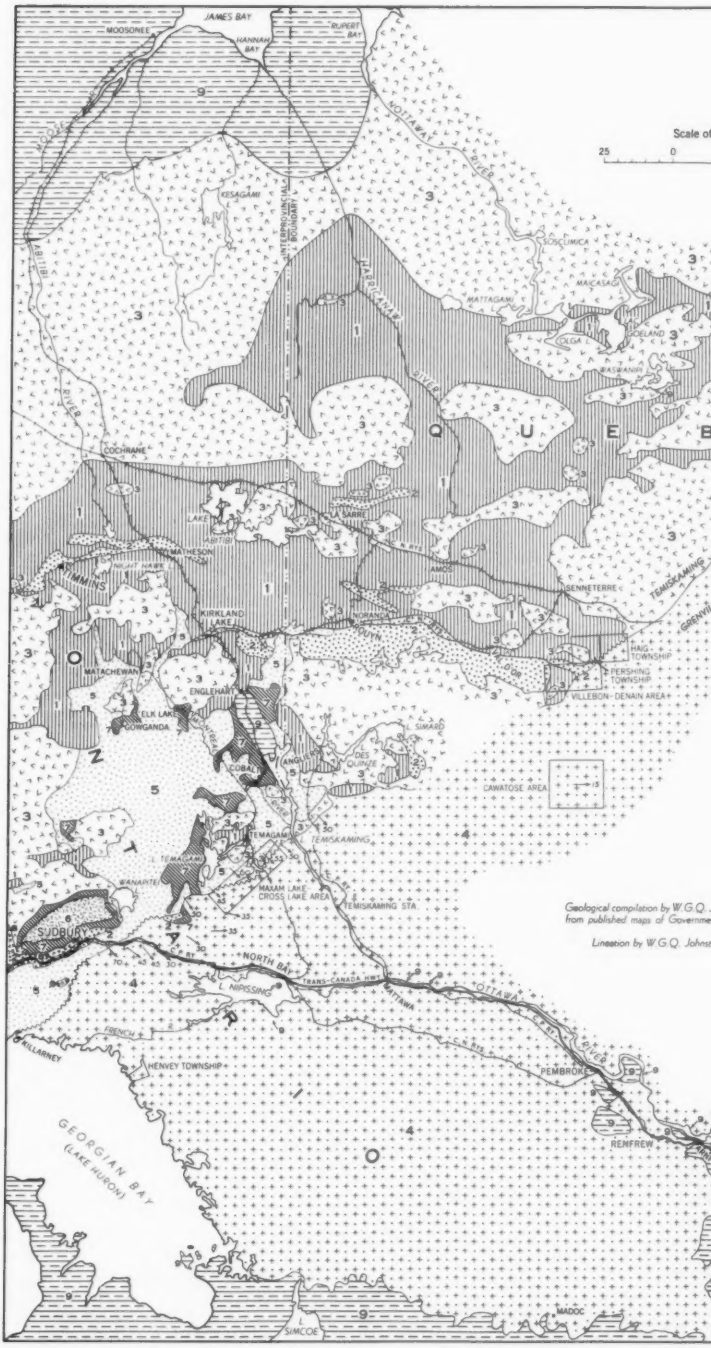
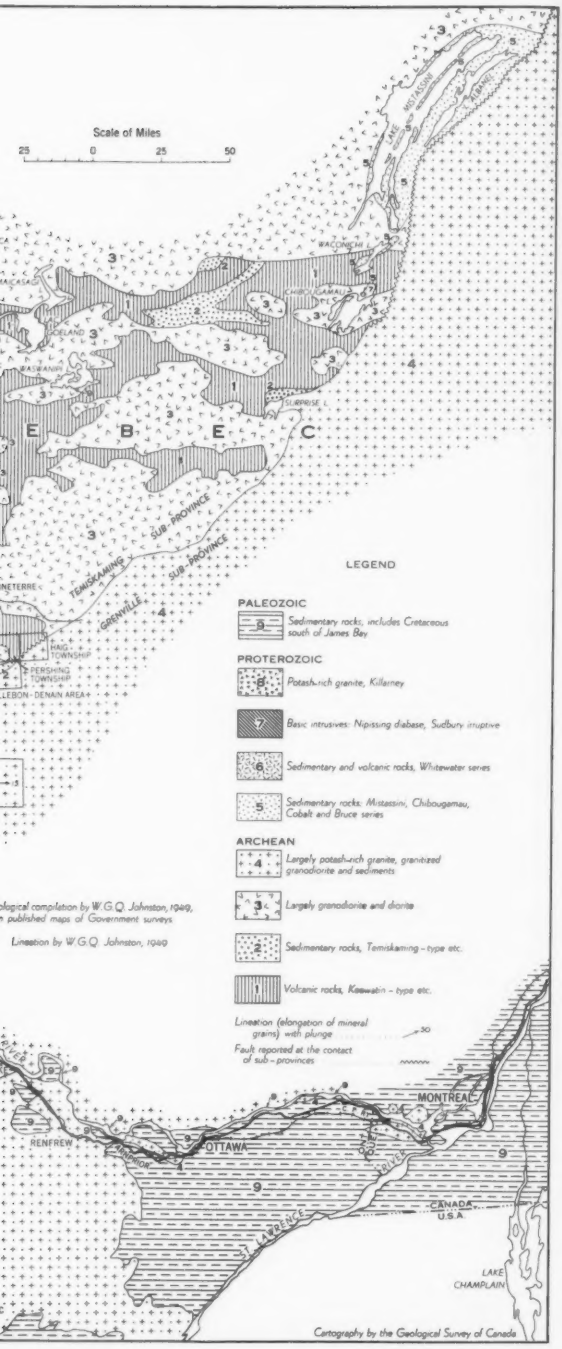



TEMISKAMING-GRENVILLE CONTA

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# GEOLOGY OF THE TEMISKAMING-GRENVILLE CONTACT SOUTHEAST OF LAKE TEMAGAMI, NORTHERN ONTARIO, CANADA

By W. G. Q. Johnston

## Abstract

The area studied lies southeast of Lake Temagami in Northern Ontario and comprises 30 miles of the contact between the Temiskaming and Grenville sub-provinces, two divisions of the Canadian Shield. The age relationship between these sub-provinces has long been a problem, and the present work was undertaken to obtain further information toward a solution. The Grenville sub-province, in the southeast part of the area, is underlain by gneisses invaded by granite. In the area studied, the Temiskaming sub-province consists of granitic rocks overlain by sedimentary rocks of the Cobalt series which are cut by sills and dikes of diabase. The work has shown that the two sub-provinces are in fault contact along a zone of northeast-striking, east-dipping faults, with displacement of the Grenville, or southeast side, upward. The amount of displacement at these faults ranks them with the major tectonic features of the earth's crust. Movement has taken place many times, the latest in late Precambrian. This movement produced a lineation in all rocks of the area and especially in the Grenville sub-province. A second set of faults strikes northwest parallel both to the movement on the earlier northeast-striking faults and to a prominent set of vertical joints. Throughout the area, northeast-striking schistosity and foliation dip southeast, and the dip flattens progressively south-eastward. Metamorphism increases in grade in the same direction. In granitic rocks of the Temiskaming sub-province the biotite isograd and zones of increasing metamorphism parallel the contact between the sub-provinces. The gneisses in the Grenville sub-province immediately east of the contact are similar in chemical composition to the granites of the Temiskaming sub-province and are considered their metamorphosed equivalent. These gneisses are cut by younger granitic rocks that are high in potash that probably originated, in part at least, by granitization. Both these granites are pre-Cobalt and probably post-Temiskaming.

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## Statement of Problem

The Canadian Shield has been divided into sub-provinces, each of which consists of lithologically and structurally similar rock formations. Although it is generally assumed that the Archean sedimentary and volcanic rocks in each sub-province are of approximately the same age or ages, knowledge of the geology as a whole is as yet incomplete.

Perhaps even more difficult to determine than the relative ages of areas of sedimentary and volcanic rocks with vast areas of intervening younger granite in each sub-province, is the relative ages of the sub-provinces themselves. This study deals with the problem of the age relation of the Temiskaming and Grenville sub-provinces (Pl. 1).

The Archean nonintrusive rocks of the Temiskaming sub-province are generally considered of two ages. The older, Keewatin type consists of volcanic rocks with minor quantities of sedimentary rocks; it is overlain (unconformably in some areas) by the Temiskaming series composed largely of conglomerate and graywacke. Both Keewatin and Temiskaming rocks have been folded, intruded by granite, and have suffered low-grade metamorphism. However, original structures, such as pillows in lavas and depositional features in the sediments, were preserved. In some areas in the Temiskaming sub-province a second series of Archean sediments is present. In the Noranda area this series is unconformably overlain by

the Temiskaming conglomerate (Wilson, 1943) and has been called the Pontiac series by Wilson (1913, p. 10). In contrast to the conglomerate and graywacke of the Temiskaming series, south of Noranda the Pontiac series contains argillaceous rocks and minor graphitic shales, and regional metamorphism has produced zones in the Pontiac marked by garnet, staurolite, cyanite, and sillimanite characteristic of those found in argillaceous (pelitic) rocks (Johnston, 1952, unpublished information). Consequently the Pontiac series should not be included with or called Temiskaming type.

In the Sudbury area, the early Precambrian rocks contain much quartzite and less lava than in the northern part of the Temiskaming sub-province. At Sudbury and along the north shore of Lake Huron the lower Huronian (Bruce series) is present as well as the Cobalt series (Upper? Huronian). The Cobalt series is the only Huronian between Sudbury and Lake Mistassini (Pl. 1) where the Mistassini series occurs.

The Grenville sub-province is characterized by gneisses and highly metamorphosed sediments. Unlike the early Precambrian graywackes of the Temiskaming sub-province, the Grenville sedimentary rocks of the eastern part of Southern Ontario and adjoining parts of Quebec were originally pure sandstones, shales, and limestones, apparently of marine origin. In this respect they resemble the late Precambrian sediments of the Temiskaming sub-province such as the Bruce and Mistassini series and to some extent the early Precambrian Pontiac series.

Few age determinations are available. The most reliable are for minerals from pegmatites younger than the rocks in which they occur. Sixteen analyses of uraninite from southeastern Ontario and adjoining parts of Quebec (Grenville sub-province) range from 985 to 1195 million years and average 1080 million years (Ellsworth, 1932, p. 105–106). Three analyses of uraninite from Henvey township (western extremity of the Grenville sub-province) in an area mapped as Killarney granite by Quirke and Collins (1930, Pl. 1) gave 835, 792, and 792 million years (Ellsworth, 1932, p. 105–106). Age determinations from Keweenawan (?) pitchblende near Lake Superior north of Sault Ste. Marie gave ages from 1310 to 1060 million years (C. B. Collins, A. H. Lang, S. C. Robinson, and R. M. Farquhar, 1952, p. 23). These determinations suggest the Grenville sub-province is late Precambrian. From the helium method Hurley (1949, p. 50) estimates the age of some Archean rocks in the Temiskaming sub-province to be over 2000 million years.

The foregoing is a brief and over-simplified summary. Complete accounts have been given by W. H. Collins (1925), M. E. Wilson (1939), Cooke (1946), Gunning and Ambrose (1939), Quirke and Collins (1930), and Dresser and Denis (1944). Large-scale maps of these areas are the Lake Huron Sheet (W. H. Collins, 1933) and Southern Quebec, west sheet (Geological Survey Canada, 1942).

### Importance of Problem

In the Canadian Shield almost all detailed work has been done where mineral deposits have been discovered, and many large intervening areas remain unmapped. Many important geological problems outside the mapped areas, such as the present problem, must be solved before a proper perspective of the geology of the Canadian Shield can be obtained.

Major fault zones and associated structures have been recognized as important in localizing ore deposits. Miller (1905, p. 33–37) pointed out the northwest and northeast drainage pattern in the important silver producing area at Cobalt, and in outlying areas. Mining at Cobalt has shown (Knight, 1922, p. 12–17) that northeast-trending faults were loci for mineral deposits and channels for ore solutions. In the same way northwest-trending faults are important at Cobalt. These faults increase in number and intensity southeastward, reaching their maximum development at the contact of the Temiskaming and Grenville sub-provinces, where a study of them should provide useful information as to their origin and relations.

### Important Contributions to Evolution of Problem

In 1887–1888, 1892, and 1895, Barlow (1907, p. 7) mapped a rectangular area extending north from Lake Nipissing and Mattawa to

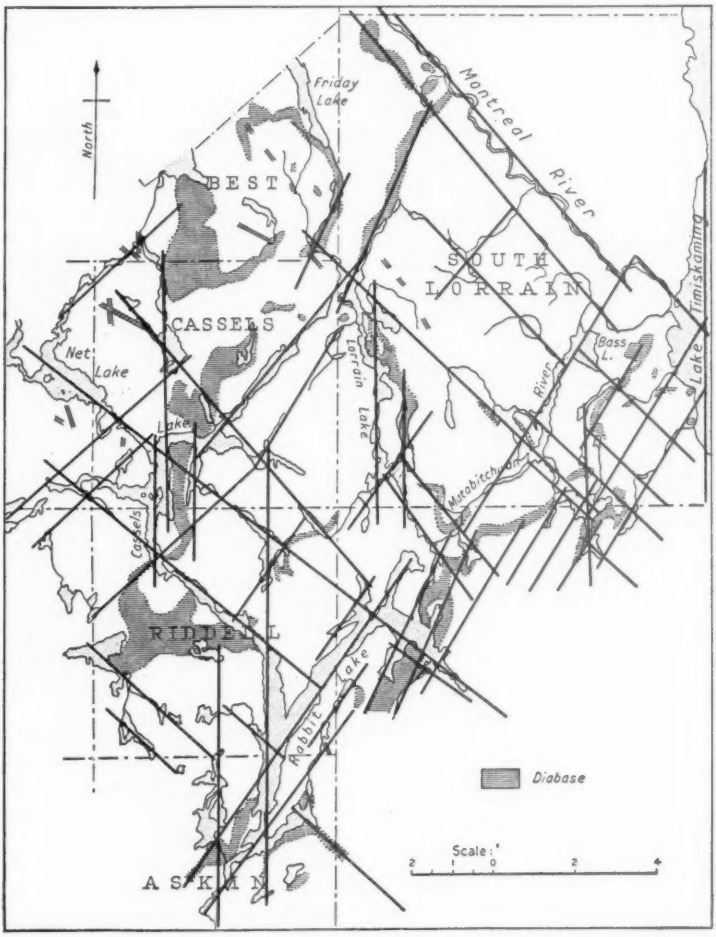


FIGURE 1.—LINES OF REGIONAL DISTURBANCE MATABITCHUAN AREA

From Todd (1925), showing fault pattern and increase in number of northeast-trending faults immediately west of sub-province contact.

the head of Lake Temiskaming (Pl. 1). Barlow noted that the gneisses (rocks in what is here called the Grenville sub-province) closely resembled the Grenville rocks of southeastern Ontario. To Barlow the high silica and alkalies and low alumina suggested magmatic origin. Their banding originated from differentiation and deformation. Barlow (1907, p. 59, 65) first reported the uniform south and southeast dip of the gneisses and the fact that their northeast strike parallels the Temiskaming-Grenville contact. He attributed this to northwest movement, during intrusion, against the rocks lying to the northwest (Temiskaming sub-province) that acted as a buttress. Barlow believed the Grenville-type[1] gneisses were younger than the Cobalt series and that the associated granite was intrusive into the Cobalt series. According to Barlow most of the gneisses

[1] Grenville-type here indicates gneisses typical of the Grenville sub-province and not the gneisses with an associated limestone which occur in type Grenville areas.

form domes. The larger domes, more than a mile in diameter, have flat-lying gneisses at the center and are much more symmetrical than the small ones. He attributed their formation to upheaval at definite lines or centers.

In 1924, E. W. Todd (1925, p. 1) mapped the Matabitchuan area (Fig. 1). Todd's most important findings were: (1) the eastern limit of the Cobalt series is determined by faults east of which the upfaulted Cobalt series was eroded, and (2) three or more generations of granite are present, all older than the Cobalt series.

Quirke and Collins (1930) studied the contact of the sub-provinces north of Killarney on Georgian Bay (Pl. 1). They found the Killarney granite (post-Cobalt age) intrusive into the Huronian sediments on the Temiskaming sub-province side of the contact and traced what they considered this same granite, containing fragments resembling metamorphosed Huronian, eastward into the Grenville sub-province. This, coupled with the fact that a thickness of over 5 miles of Huronian sediments extending some 150 miles to the west suddenly disappeared at this contact, strongly suggested either granitization or erosion. Quirke and Collins concluded the Grenville sub-province represented metamorphosed and granitized Huronian rocks.

One weakness in this hypothesis is the presence of a great fault (first described by Quirke and Collins, 1930, p. 32) at the contact between the sub-provinces. The sedimentary inclusions in the granite are much more highly metamorphosed southeast of the fault than are the Huronian sediments immediately northwest of it. W. H. Collins (1936, Pl. 2), in later recognizing the importance of this fault, appears to have abandoned the hypothesis of a Huronian age for the sediments of the Grenville sub-province. Cooke (1946, p. 16) also argued against the Grenville rocks being Huronian.

East-west folding in the Huronian rocks extends 150 miles to the west along the north shore of Lake Huron and is abruptly transected by northeast-trending folding just east of the great fault.

From observations along the Temiskaming-Grenville contact in northwest Quebec, Norman (1936, p. 120) was the first to point out that a continuous zone of faulting from Lake Chibougamau to Georgian Bay might mark the contact of the sub-provinces (Pl. 1). Norman (1940, p. 522; 1946, p. 132) dated this faulting and mountain-building as late Precambrian because: (1) East-west folding in the Keewatin-type rocks (Archean) is truncated by later northeast folding and faulting at the contact of the Grenville and Temiskaming sub-provinces, as pointed out by A. M. Bell (1932, p. 70–73) who wrote:

> 'Towards the south, metamorphism of the beds has been more intense and they grade into gneisses under the influence of deep burial and lit-par-lit granitic injections. The northeast folding appears to have been superimposed on the east-west folding which in places has been dragged sharply into the northeast direction."

(2) In the vicinity of Lake Chibougamau this faulting produced zones of sericite schist in late Precambrian sediments, which Norman correlated with the Cobalt series on the basis of: (a) marked lithological similarities, such as pebble-bearing, varved graywackes; (b) similar relationships to Archean rocks. And (3) thrusting of the granite over late Precambrian sediments occurred near Lake Mistassini to the north of Lake Chibougamau (Norman, 1940, p. 521).

In Haig township (Pl. 1) Norman (1947, p. 10) postulates a fault between relatively unmetamorphosed, easterly trending flows and sediments on the northwest side and northeasterly trending biotite and hornblende gneisses on the southeast side. In adjoining Pershing township, this fault is marked by schistose graywacke and mylonitized granite.

Norman (1947, p. 11) believes the gneisses southeast of the fault in Pershing and Haig townships developed in Archean time and may represent a gradual transition from normal graywacke (Temiskaming-type sediments) southeastward into garnetiferous biotite gneisses. This correlation, one of the earliest advanced for the relationship between the two sub-provinces, has also been put forward by Lowther (1936, p. 47) in Villebon and Denain township (Pl. 1).

The zone of faulting extending to Lake Mistassini (Pl. 1) studied by Norman and others may extend northeastward to the

Labrador coast. Cooke (1947, p. 21) states that thrust faults bound the late Precambrian rocks of the Labrador trough on the south. Still farther east, at Seal Lake, late Precambrian sedimentary rocks are cut off to the south by thrust faults (Roscoe and Fahrig, 1952, oral communication). Hence in the Labrador peninsula Grenville-type rocks lie on both sides of the assumed fault zone.

## Scope of Present Work

The Maxam Lake-Cross Lake area (Pl. 1) is roughly 30 miles long by 10 miles wide and has a number of advantages over other areas embracing this same geological problem. The foremost of these is the abundance of outcrops, even though the area is heavily wooded and many of the outcrops are moss covered. Many good exposures occur around the lakes and in rock cuts along the highways and railroad.

Topographically, the Temiskaming sub-province portion of the area is much more rugged than the Grenville sub-province portion, probably because the former is underlain by rocks of variable resistance to erosion. In contrast the Grenville sub-province is flat, because of its thoroughly recrystallized rocks of about equal resistance to erosion.

Faults control the courses of streams and location of lakes (Fig. 1). Many lakes lie where faults unite or cross.

Drift is sparse. Much fine sand occurs east of Red Cedar Lake. Glacial striations indicate ice movement of 190° throughout the area. Less prominent striations at 155° occur in a few places. Chatter marks are common in areas of Lorrain arkose.

The writer working alone did 7 months field work from 1945 to 1949, mainly during 1948 and 1949. Laboratory investigations included the study of 300 thin sections, preparation of 20 fabric diagrams, and 30 statistical diagrams of structures.

## Map Units

### *General*

Early Precambrian volcanic and sedimentary rocks are not present in the Temiskaming sub-province portion of the area which is a disadvantage. However, the Cobalt series and two pre-Cobalt granites occur. The Grenville sub-province portion is underlain by gneisses and granite.

Events in the geological history have been included in the map legend.

### *Granite 1*

The designations granite 1 and granite 2 are used rather than conventional names such as Algoman which imply a specific age. In this work granite is used loosely and includes granitic rocks in general. Granite 1 is confined to the Temiskaming sub-province. It is coarse-grained, massive, and uniform except for occasional banded dark-green fragments of what appears to be recrystallized lava. It is grayish green due to epidote. The chief constituents are plagioclase, quartz, and hornblende. Accordingly the rock is quartz diorite or granodiorite. The chief variation is in the amount of quartz. Biotite and microcline each constitute less than 5 per cent. Accessory minerals are sphene, apatite, pyrite, magnetite, fluorite, and zircon. The plagioclase is considerably altered to epidote-albite white mica, and the hornblende to chlorite and epidote. Except for a few areas representing inclusions of lavas this rock is the oldest in the Temiskaming sub-province.

### *Granite 2*

Granite 2 occurs in the Grenville sub-province. In the Temiskaming sub-province a granite is present, which in every respect, resembles granite 2 in the Grenville sub-province and thus is believed to be the same granite. Southeastward from the contact of the sub-provinces massive granite gives way to a gneissic variety. South of the railroad there is an exceptionally wide band of massive granite (disregarding areas of shearing). Southeast of the sub-province contact it extends eastward to Milne Lake and the eastern extremity of Red Water Lake.

Granite 2 is pink, aplitic, and medium- to fine-grained. It is composed of feldspars and 10–25 per cent quartz. Biotite is present in amounts up to 5 per cent. Sphene, much of it

polysynthetically twinned where deformed, is present in amounts up to 5 per cent. Other minerals are magnetite, pyrite, hematite, apatite, carbonate, epidote, and allanite. $An_{25}$ at the contact of granite 2 and the older gneisses. The potash feldspar is usually microcline, occasionally untwinned. Grains composed of both twinned and untwinned varieties give

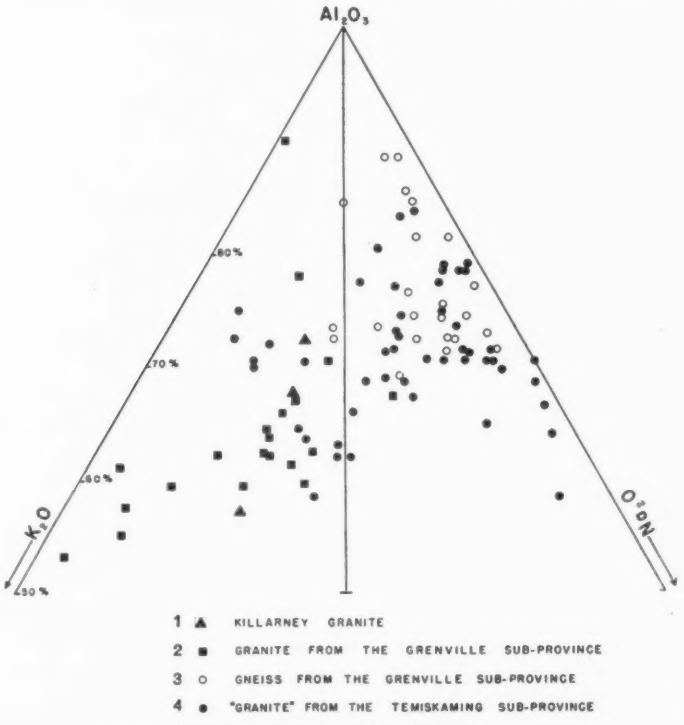


FIGURE 2.—PLOT OF 105 ANALYSES

Showing difference in $K_2O$ content of granitic rocks in Grenville and Temiskaming sub-provinces and similarity of Temiskaming sub-province granites and Grenville-type gneisses.

1—Killarney granite; 3 chemical analyses from Temiskaming sub-province; 2—Granite from western part of Grenville sub-province (granite 2); 18 chemical analyses, 15 reported as Killarney granite within Grenville sub-province; 11 Rosiwal analyses, 4 from Maxam Lake-Cross Lake area; 3—Gneisses from western part of Grenville sub-province; 3 chemical analyses; 19 Rosiwal analyses; 4—Granite from Temiskaming sub-province (Algoman, granite 1, and others); 46 chemical analyses; 4 Rosiwal analyses from Maxam Lake-Cross Lake area.

References: Barlow (1907); Gussow (1937); Jones (1930); Quirke and Collins (1930); Wahl and Osborne (1950); remainder largely from reports of Ont. Dept. of Mines.

Grains of allanite are usually amorphous and not enclosed in epidote as in the gneisses. In contrast to granite 1 which is highly and uniformly altered, granite 2 is fresh, except along zones of dislocation where the feldspars are altered to white mica. Minerals in granite 2 are fresh and clear. Potash feldspar is 2–4 times as abundant as plagioclase. The plagioclase ranges from $An_{12}$, the normal composition, to the same extinction angle. Cleavage is always better developed in orthoclase than in microcline. Granite 2 is much higher in $K_2O$ than granite 1 (Fig. 2). Thus these granites belong to separate petrographic provinces. The high $K_2O$ content and presence of allanite in granite 2, in both sub-provinces, suggest this granite is similar to the granites of southeastern Ontario.

This high $K_2O$ content of granite in the Grenville sub-province has been noted by Barlow and Young (1907) to the west and Henderson (1936, p. 18) just east of Lake Temiskaming. Gussow (1937) made a compositional study of granite bodies in the Northwestern Quebec gold belt and vicinity and found they were higher in soda than potash, and some exceptionally so. The high potash content of granite 2 in the area under discussion makes it similar to the Killarney granite (post-Huronian) as emphasized by Quirke and Collins (1930, p. 102).

The Cobalt sediments overlie granite 2 in the Temiskaming sub-province. Granite 2 irregularly invaded granite 1, even to some extent along joints; this and the moss cover and the widespread and intense shearing make it extremely difficult to differentiate the two.

### *Cross Lake Series*

This name has been given to fine-grained, bedded sediments and minor, associated (?) volcanic rocks underlying areas in the Temiskaming sub-province around Cross Lake and the small lake to the southwest. Barlow (1907) mapped these rocks as Cobalt series. They have been recrystallized to biotite, white mica (forming nodules on the weathered surface), and untwinned plagioclase. These rocks are overlain unconformably by the Gowganda formation which shows no recrystallization. The Cross Lake series is cut by granite 2, but was not found in contact with granite 1. South of the area of Cross Lake series in the Grenville sub-province are similar although coarser-grained, more thoroughly recrystallized rocks. Hence, some of the Grenville-type gneisses may be the metamorphosed equivalent of the Cross Lake series.

### *Grenville-type Gneisses*

These rocks underlie large areas of the Grenville sub-province. No rock positively identified as volcanic or sedimentary was found in this group. The predominant type is a coarse-grained gray- to brown-weathering rock composed entirely of plagioclase, biotite, and quartz. Most of the amphibole in these gneisses is in rock forming boudins. Although these rocks appear variable in the field, microscopically they are remarkably similar in composition and mineralogy. Their most striking feature is a foliation marked by layers rich in biotite, in places folded and crumpled. These rocks are the oldest rocks in the Grenville sub-province and have been invaded by granite 2.

### *Basic Dikes*

Throughout the area, especially around Jumping Caribou Lake, narrow, green, fine grained, sheared dikes occur. These cut both granite 1 and granite 2 but were not found in contact with the Cobalt series.

### *Marten River Granite*

This dike-like body of granite trends north eastward from the southeast extremity of Red Cedar Lake. It is a coarse-grained augen rock composed of microcline, biotite, and quartz and is markedly different from and younger than granite 2.

### *Gowganda Formation*

The Huronian is represented in this area by the Cobalt series (upper? Huronian) which consists of two formations—the Gowganda overlain by the Lorrain. The eastern limit of a large area of Cobalt series to the west lies in the map area.

Unlike other areas where this formation is flat-lying, folding, faulting, and the irregular terrain have exposed excellent sections. Ten to 40 feet of conglomerate with a high proportion of boulders to graywacke matrix occurs at the base. The boulders are angular at the base and in many places have moved only a few inches. Around Reuben Lake, erosional remnants of conglomerate remain as unconnected joint and crevice fillings in the basement granite. Well-formed pebble dents are common and indicate tops of steeply dipping beds. Boulders up to 5 feet in diameter have dented underlying beds. Mud cracks and flow casts (Shrock, 1948, p. 156) are common.

On the east shore of Ingall Lake, north of the portage to Angus Lake, the matrix of the conglomerate is lighter in color and coarser adjacent to the cobbles. These sandy parts of

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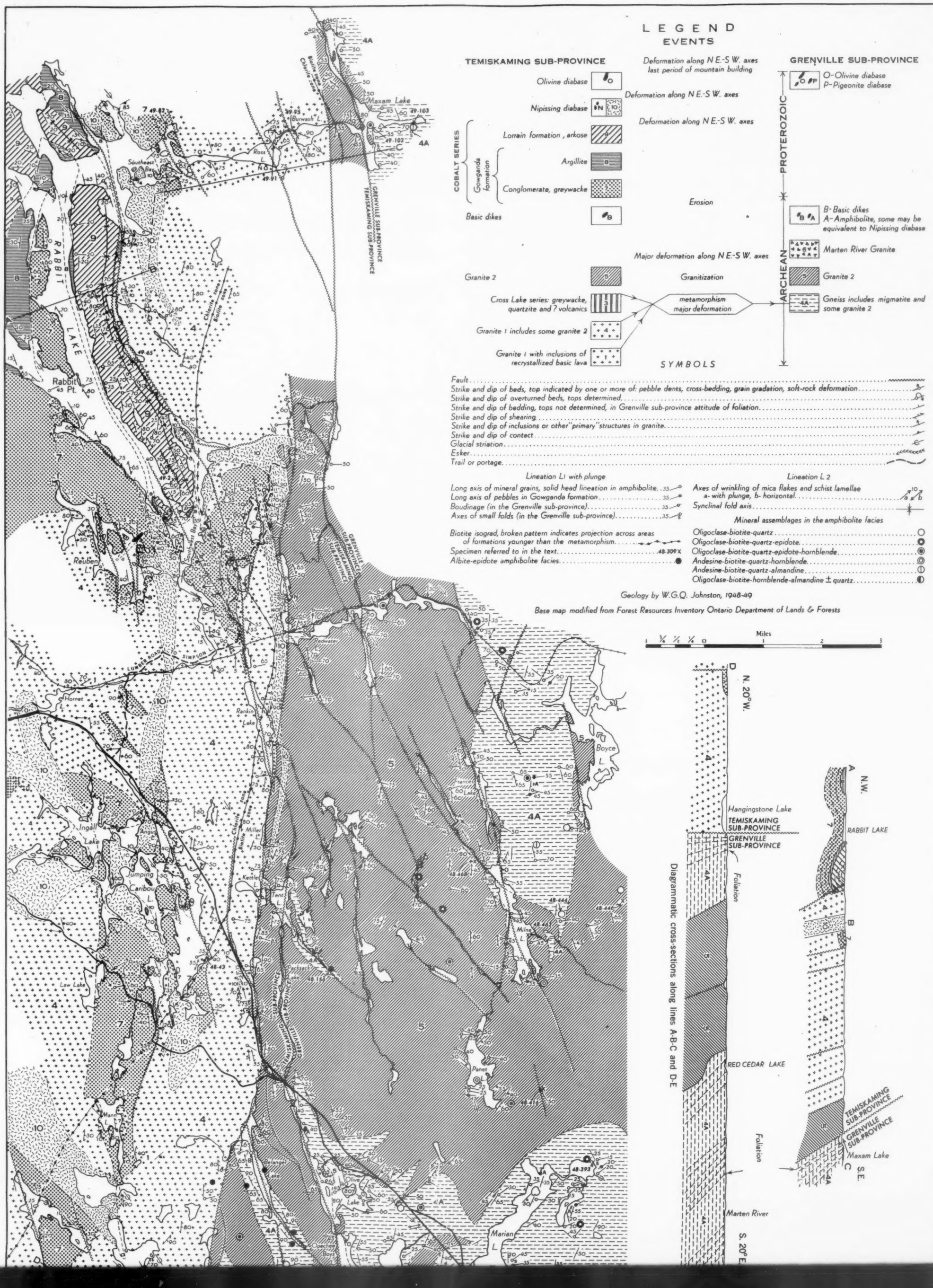
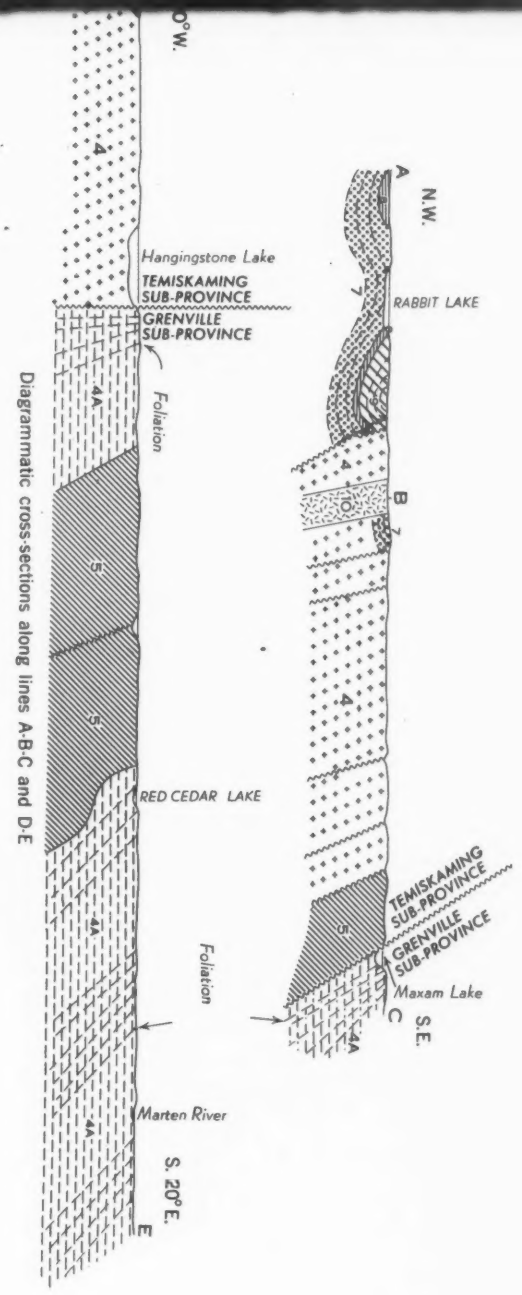
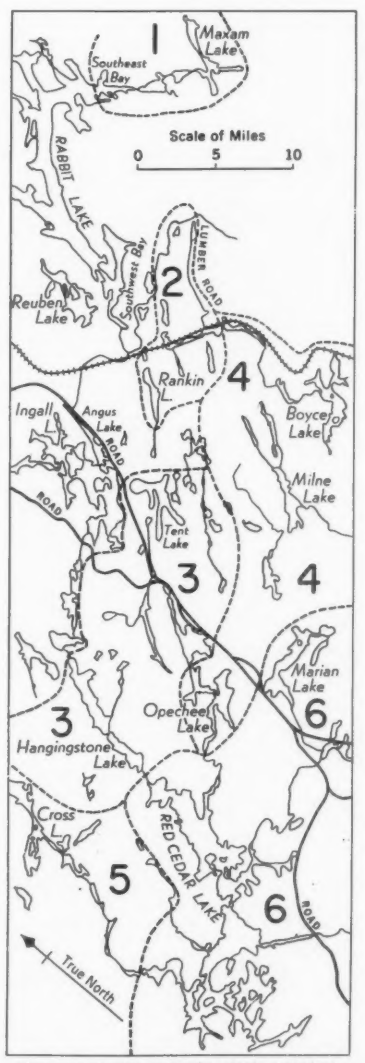

LEGEND
EVENTS
TEMISKAMING SUB-PROVINCE
Deformation along N.E.-S.W. axes last period of mountain building
Olivine diabase
Deformation along N.E.-S.W. axes
Nipissing diabase
Deformation along N.E.-S.W. axes
COBALT SERIES
Gowganda formation
Lorrain formation, arkose
Argillite
Conglomerate, greywacke
Erosion
Basic dikes
Major deformation along N.E.-S.W. axes
Granitization
Granite 2
Cross Lake series: greywacke, quartzite and ? volcanics
metamorphism major deformation
Granite 1 includes some granite 2
Granite 1 with inclusions of recrystallized basic lava
GRENVILLE SUB-PROVINCE
O-Olivine diabase
P-Pigeonite diabase
PROTEROZOIC
B-Basic dikes
A-Amphibolite, some may be equivalent to Nipissing diabase
Marten River Granite
Granite 2
Gneiss includes migmatite and some granite 2
ARCHEAN
SYMBOLS
Fault
Strike and dip of beds, top indicated by one or more of: pebble dents, cross-bedding, grain gradation, soft-rock deformation
Strike and dip of overturned beds, tops determined
Strike and dip of bedding, tops not determined, in Grenville sub-province attitude of foliation
Strike and dip of shearing
Strike and dip of inclusions or other "primary" structures in granite
Strike and dip of contact
Glacial striation
Esker
Trail or portage
Lineation L1 with plunge
Long axis of mineral grains, solid head lineation in amphibolite
Long axis of pebbles in Gowganda formation
Boudinage (in the Grenville sub-province)
Axes of small folds (in the Grenville sub-province)
Lineation L2
Axes of wrinkling of mica flakes and schist lamellae a- with plunge, b- horizontal
Synclinal fold axis
Biotite isograd, broken pattern indicates projection across areas of formations younger than the metamorphism
Specimen referred to in the text
Albite-epidote amphibolite facies
Mineral assemblages in the amphibolite facies
Oligoclase-biotite-quartz
Oligoclase-biotite-quartz-epidote
Oligoclase-biotite-quartz-epidote-hornblende
Andesine-biotite-quartz-hornblende
Andesine-biotite-quartz-almandine
Oligoclase-biotite-hornblende-almandine ± quartz
Geology by W.G.Q. Johnston, 1948-49
Base map modified from Forest Resources Inventory Ontario Department of Lands & Forests
Miles
Diagrammatic cross-sections along lines A-B-C and D-E
N. 20° W.
N.W.
S.E.
S. 20° E.
Hangingstone Lake
TEMISKAMING SUB-PROVINCE
GRENVILLE SUB-PROVINCE
RABBIT LAKE
RED CEDAR LAKE
Maxam Lake
Marten River
Foliation
Maxam Lake
Burwash L.
Ross L.
Southeast Bay
Rabbit Pt.
Reuben L.
Redwater
Hornet L.
Rankin Lake
Boyce L.
Frances Lake
Ingall Lake
Angus L.
Jumping Caribou L.
Kettle L.
Tent L.
Miller L.
Law Lake
Mann L.
Belanger Lake
Olive Lake
Milne L.
Panet
Marian L.
Peshogan Bay
Lumber Road
Simpson L.
Rabbit Lake
Chlorite zone
Biotite zone

Diagrammatic cross-sections along lines A-B-C and D-E




Key map showing areas for which statistical structure-diagrams were prepared.



Cartography by the Geological Survey of Canada

GEOLOGICAL MAP OF MAXIM LAKE-CROSS LAKE AREA

the matrix form strikingly concentric envelopes about 1 inch thick around and adjacent to the cobbles or less than 1 inch from them. Thin sections of these envelopes indicate the quartz, chlorite, and an authigenic? amphibole are coarser in grain than the chlorite and quartz in the rest of the matrix.

One outstanding feature of the Gowganda formation is the widespread occurrence of soft-rock deformation, best developed in the lower horizons and always involving more than one bed. Crumpling, small-scale folding, and faulting were observed but no brecciation. The soft-rock deformation seems to have been confined to the surface beds, as many of these are truncated. No soft-rock deformation definitely assignable to overriding by ice was observed. A soft-rock foliation has developed in some places, apparently marking surfaces along which movement took place prior to consolidation. These surfaces are curved and even corrugated. They show no consistency of strike or obvious relationship to regional schistosity.

The conglomerate, at the base of the Gowganda formation gives way upward to a rock in which the graywacke matrix greatly predominates over the fragments in which lenses of fine-grained pink arkose occur. In neither the conglomerate nor the graywacke is bedding consistently present. Bedding is exceptionally well graded. Where careful scrutiny fails to reveal bedding, joints are present which, if traced, are parallel to discernible bedding.

East of Southeast Bay of Rabbit Lake the thickness (estimated) is 2300 feet.

In some localities the Gowganda formation was deposited on a surface of considerable relief (Hopper, 1942, p. 389; Ambrose and Ferguson, 1945, p. 13). This surface was extremely irregular and perhaps even mountainous. In the map area dips of beds are gentle, while contacts between the underlying granite and the Gowganda formation dip steeply. The widespread soft-rock deformation may be the result of sliding during adjustment to the extremely irregular underlying surface.

No criteria diagnostic of a glacial origin for the Gowganda formation were found. The regolith beneath the Gowganda formation of this area seems difficult to reconcile with a glacial origin.

### *Lorrain Formation*

In the map area this formation occurs only around Rabbit Lake where it is about 1700 feet thick. The normal Lorrain is a coarse-grained arkose. The high feldspar content differs from compositions elsewhere. To a depth of half an inch the Lorrain is weathered gray; below this it is a characteristic pale green. Lenses of well-rounded granite pebbles up to an inch in diameter are not uncommon. The base of the southern part of the Lorrain is a fine-grained pink arkose. Current ripples were observed, and well-developed cross-bedding indicates a source of sediment to the north.

Magnetite is the most abundant heavy mineral. Angular to well-rounded zircons are widespread. They average 0.25 mm, but some are more than 0.5 mm long. The larger zircons are brown, and many have adhering quartz; the smaller ones are mauve and have more inclusions. Other heavy minerals in approximate order of abundance are: amphibole, garnet (often with gas cavities), tourmaline, leucoxene, muscovite, carbonate, epidote, apatite, sphene, vesuvianite (?), fluorite, and corundum.

The Gowganda formation was definitely derived from the Temiskaming sub-province as evidenced by the abundance of lava fragments and granites typical of the Temiskaming sub-province. Not a single boulder or pebble of gneiss characteristic of the Grenville sub-province was found in the Gowganda formation. However, the heavy-mineral suite of the Lorrain indicates it was, in part at least, derived from the Grenville sub-province. The occurrence of amphibole fragments with adhering garnets indicates that an area similar to the Grenville sub-province was supplying sediment when the Lorrain formation was deposited. This suggests the possibility of an important break between the Gowganda and the Lorrain as has been described east of Lake Temiskaming by Henderson (1936, p. 24). However, brief examination suggests a gradation between the two formations in the present map area. Apparently the Gowganda argillite becomes more quartzose, thicker bedded, and dark green and this dark-green rock grades upward into the more massive, lighter-colored, coarser-grained Lorrain.

### *Nipissing Diabase*

Marked compositional and textural variations in this rock suggest more frequent and larger earth movements during its intrusion in this area than in areas to the west where it tends to be more uniform. Also chilled contacts within the bodies along with sharp variations in composition suggest multiple and composite intrusions.

### *Amphibolites of Grenville Sub-province*

These rocks are probably not all the same age. However, the writer believes they were the same general rock type (diabase) before metamorphism. They occur in sills, usually less than 50 feet wide. They are green-weathering, medium-grained rocks; some contain large feldspar crystals, probably phenocrysts of the original porphyritic diabase.

They pose another difficult and interesting set of problems. The youngest diabase—the olivine diabase, occurs in both sub-provinces. The abundance of Nipissing diabase in the Temiskaming sub-province, close to the contact of the sub-provinces, suggests it should also occur in the Grenville sub-province. If so, it must occur as amphibolite. Just east of the north portion of Olive Lake are amphibolites megascopically similar to the Nipissing diabase.

### *Olivine Diabase*

This rock, the youngest rock in either sub-province, aids in establishing contemporaneity of events on either side of the contact of the sub-provinces. Just west of the area in the Temiskaming sub-province, along No. 11 Highway, it cuts the Nipissing diabase.

In the Temiskaming sub-province the olivine diabase is weathered to a brown mantle up to a foot thick. Deep weathering distinguishes it from the Nipissing diabase. The unweathered rock is gray, has a more pronounced ophitic texture, and is fresher than the Nipissing diabase. It is also commonly porphyritic (plagioclase phenocrysts), a feature not found in the Nipissing diabase.

Although recognizable by freshness and strongly developed ophitic texture, there are some differences in the olivine diabase dikes in the Grenville sub-province. The feldspars are dark brownish purple, owing to inclusions which appear as dust under the highest magnification. Large feldspar phenocrysts are not present in the olivine diabase in the Grenville sub-province.

These olivine diabase dikes in the Grenville sub-province are similar to the large olivine-bearing diabase intrusion in Flett and Angus townships, about 2 miles east of the Maxam Lake-Cross Lake area. Hurst (1931, p. 105) who studied this occurrence was unable to determine the age relationship of the diabase and enclosing gneiss. A garnet-bearing amphibolite occurs at the diabase-gneiss contact. The amphibolite is in sharp contact with the gneiss and is similar to amphibolite zones in olivine diabase dikes, which in places exhibit chilled contacts against similar enclosing gneisses in the Maxam Lake-Cross Lake area. Hence from a limited study of the Flett and Angus township diabase the writer believes it is younger than the enclosing gneiss.

### *Pigeonite-Quartz Diabase*

Only four of these dikes were identified and only one in the Temiskaming sub-province, at the railroad about half a mile south of the lumber road crossing. Megascopically the dikes are similar to the olivine diabase and are probably genetically related to it. They contain quartz, and the pyroxene is uniaxial pigeonite. Indices of pigeonite from specimen 48–326 (north portion Red Cedar Lake) are 1.677 and 1.702, $Z \wedge C = 30°$. The plagioclase is near $An_{45}$, the same composition as that in the olivine diabase.

## Structural Geology

### *Folds*

The gneisses dip predominantly south and southeast. Minor folds are visible in areas of good exposure.

A syncline occurs in the Gowganda formation east of Southwest Bay of Rabbit Lake. Beds are overturned more than 20° in some places on the east limb of this fold. Two cross-sections illustrating folding and attitudes of foliation are shown on Plate 2.

### *Faults*

The three main sets of faults in the area strike north, northwest, and northeast; the northeast set is dominant.

Evidence for northeast[2] faults is manifest in air photographs. In places, for widths of over a mile at the Grenville front fault zone the rocks are heavily sheared. However at Hangingstone and Cross lakes (Pl. 2, areas 3, 5) the fault is sharply defined, and rocks of the Grenville sub-province occur about 200 feet from undeformed rocks of the Temiskaming sub-province. In the vicinity of Tent Lake (area 3) there are two well-defined faults, one at Kettle Lake, the other at Packsack Lake and Tent Lake. To the north, the best-defined of these is the fault extending through Miller and Rankin lakes. The granite along the lumber road east of Rabbit Lake is heavily sheared. Strong shearing occurs between Burwash and Ross lakes (area 1). The main part of Maxam Lake follows a strong fault.

At the Grenville front fault there is a profound lithologic change. Where granite 2 and not Grenville-type gneisses occurs in contact with the Temiskaming sub-province, the contact of the sub-provinces has been indicated where the character of granite 2 and its inclusions are similar to those in the Grenville sub-province. This contact between the sub-provinces is a complicated and irregular zone of faulting striking northeastward along which successive north segments lie to the east (Pl. 2).

Schistosity in the vicinity of the Grenville front fault zone dips 60° E. to vertical. Where schistosity symbols are not shown on the map lineation $L_1$ gives a close approximation to the direction and angle of dip of schistosity.

The displacement along the Grenville front fault zone (east side upward) was large, probably miles. Some conception of the magnitude of this displacement can be obtained from the observation that 4000 feet of Cobalt sediments in the north part of the area has been upfaulted and eroded along faults lying east of, and showing only slight deformation compared to, the faults at the main zone of faulting.

Strong zones of faulting occur within the Grenville sub-province, such as at Milne Lake (Pl. 2, area 4) and the two small lakes to the north.

Strong zones of faulting also occur within the Temiskaming sub-province. The fault at Rankin Lake (area 2) appears to split, and one branch cuts off the Cobalt series east of Southwest Bay, Rabbit Lake. The granite on the small granite island north of the largest island in this part of Rabbit Lake is heavily sheared, as is the arkose to the north along its eastern limit. This is especially noticable at the two small lakes east of Rabbit Point. Here a 12-foot band of ground-up arkose containing vuggy quartz stringers occurs at the arkose-granite contact. However, the intensity of deformation in the arkose indicates that the fault of largest displacement trends up the long axis of the two lakes, and here, as in most places, the fault which cuts off the thickest section of the Cobalt series is not at the contact of the granite and Cobalt sediments but west of it.

Except for minor carbonate and quartz, there has been very little introduction of material along these faults during and following latest movements.

### *Schistosity*

In the Temiskaming sub-province northeast schistosity is predominant and occurs near faults with which it seems parallel. Although northeast schistosity is confined to certain areas, northeast joints are widespread in the Temiskaming sub-province.

In the Grenville sub-province schistosity tends to follow an earlier foliation. Although it parallels the foliation in strike, later schistosity often has a much steeper dip, especially at faults.

The development of schistosity in either sub-province can probably be reconstructed by observing the changes from unsheared to heavily sheared granite: (1) flattening of quartz grains by flowage to elongated flattened forms (Pl. 3, fig. 3); (2) still further flattening of quartz until the grains unite to form layers of almost pure quartz separated by layers of feldspar; (3) rupture between the quartz and

[2] For the sake of simplicity northeast-striking, eastward-dipping faults, schistosity, and joints are referred to as northeast faults, schistosity, and joints. Similarly, northwest-striking, steeply dipping faults, schistosity, and joints are referred to as northwest faults, schistosity, and joints.

feldspar layers and within the quartz; and (4) further movement reducing the rock to a schist and the individual quartz-feldspar layers losing their identity through fragmentation.

### *Foliation*

This structure (used in the sense of Darwin as quoted by Harker, 1939, p. 203) is found in the gneisses of the Grenville sub-province. It seems to represent a concentration of biotite along prerecrystallization *S*-surfaces. This separation of light and dark minerals into bands has given the gneisses a "bedded" appearance, and they might be designated paragneisses by some. Some gneisses, east of the Temiskaming-Grenville contact, lack foliation because they are less deformed.

At Red Cedar Lake and along the Temagami River the foliation and schistosity in both granite and gneiss is in the form of large curves convex westward. Due to the eastward dips the west parts of these structures crop out more prominently, and it is not known whether they form complete circles in plan. Considered in three dimensions the foliation and later parallel schistosity seem to be arranged in partial concentric cones. These appear nearly conical and hence do not seem to be the corrugation folds found in some thrust planes. Their origin might be related to the principle that failure in a homogeneous body under compression is along conical surfaces.

### *Relation of Three Directions of Fracture*

To analyse the various directions of fracture, 10 diagrams (Pl. 4, nos. 1–10), consisting of 2232 poles to *S*-surfaces (exclusive of bedding in the Cobalt series) were plotted. The areas for which these diagrams were prepared have been numbered on the key map (Pl. 2). These diagrams give considerable information: the northeast schistosity and foliation flatten progressively eastward, *i.e.* schistosity in the Lorrain Formation (Pl. 4, no. 6), dips 70° while in the most easterly studied gneisses it dips 45° (Pl. 4, no. 1).

It has long been known that three prominent directions of weakness control topography in this area. Todd's (1925, p. 28) sketch map of these in the Matabitchuan area is reproduced here (Fig. 1). The northeast lines of weakness are more prominent and more numerous to the southeast.

Strong north-south lines of weakness also appear in this figure, but the pole to *S*-surface diagrams (Pl. 4, nos. 1–10) show that relatively few north-south fractures were recorded. Field observations suggest the north-south faults (possibly parallel to some older direction of fracture) represent adjustments to differences in rates of movement between blocks moving upward and northwestward on the northeast faults. For example a strong north-south fault, at Belanger Lake (Pl. 2, area 3), terminates at the fault marking the contact of sub-provinces. Apparently another strong north-south fault through Angus Lake (Pl. 2 N.W. of area 3) and north of the map area through Cassels, Pishabo, and Rib lakes marks the west boundary of an area underlain by Lorrain Formation, indicating that relative movement was east side down. As will be explained later the degree of recrystallization on either side of the north-south fault at Belanger Lake (Pl. 2, area 3) suggests this same displacement.

North-south and northeast faults have lineations of about the same attitude. The schistosity near north-south faults is northeast in many places.

Northwest joints dipping steeply and predominantly southward are common. They are smooth, regular, continuous, uniformly spaced (usually 4 or 5 inches apart), very rarely coated with later minerals, and consistent in attitude. They cut clearly through and are never displaced by the northeast schistosity. These northwest joints are best developed and most numerous in areas of strongest northeast shearing. The pole diagrams for *S*-surfaces (Pl. 4, nos. 1–10) show that the two directions of fracture maintain a right-angle relationship in plan. Hence the northwest joints are probably tension fractures parallel to the northwestward movement, on northeast-striking, eastward-dipping schistosity. This origin is substantiated by observations around Red Cedar Lake (Pl. 2, area 6) where the two sets of fractures are not quite at right angles (Pl. 4, no. 1) but the lineation $L_1$ (parallel to movement) is parallel to the northwest set of joints (Pl. 4, no. 12).

*Direction of Tectonic Transport*

The northeast faults brought entirely different types of rocks into juxtaposition, and displacement must have been great. The following evidence strongly suggests that movement was east side upward and northwest:

(1) The entire thickness of the Cobalt series, probably in excess of 4000 feet, has been cut off on faults near Rabbit Lake, and its eastward extension has been eroded. At these faults west-facing granite cliffs stand as high as 200 feet above the Cobalt series; the strata here show no evidence of having been deposited at the foot of cliffs.

(2) Along the faults the Gowganda formation has been overturned to the west and dips under the older granite (Pl. 3, figs. 1, 2). This can also be seen on the east side of the small lakes east of Southwest Bay of Rabbit Lake. These folds with horizontal axes and overturned east limbs are the result of movement normal to their axes *i.e.*, an upward and northwestward movement on the faults. A further and much more widespread example of fold axes normal to the above movement is lineation $L_2$ consisting of horizontal wrinkling axes and ridges.

(3) A few west-dipping fractures striking parallel to the northeast schistosity may be tension cracks. All those observed indicate a relative movement of east side upward (Pl. 4, nos. 1, 2, 7, 9).

(4) The metamorphism which increases in grade from northwest to southeast suggests displacement was east side upward (Fig. 5).

(5) The prominent lineation $L_1$ usually plunges at right angles to the strike of northeast foliation and schistosity. The second lineation $L_2$, also in the plane of foliation or schistosity, consists of wrinkling of schist lamellae and small ridges with approximately horizontal axes. Thus wrinkling and fold axes are at right angles to the prominent lineation $L_1$. Movement is usually parallel with or at right angles to lineation. A consideration of the previous points favors movement parallel to $L_1$. Also, $L_1$ appears too straight and parallel to have formed by any movement other than one parallel to it.

(6) Slickensides west of the main zone of faulting, and irregularities on all fault surfaces examined, indicate the relative movement was east side up and to the northwest.

(7) The apparent increase in specific gravity of the olivine diabase dikes southeastward from the Temiskaming-Grenville contact (Fig. 3) might be interpreted as due to a progressively deeper section of these dikes being exposed at the surface. The dikes of higher specific gravity have a higher proportion of olivine and pyroxene relative to plagioclase, and contain slightly more magnetite. However these differences could be due to factors other than depth of formation.

*Hanging-Wall Versus Foot-Wall Movement on Northeast Faults*

There remains the problem of distinguishing the active block during reverse movement on the northeast faults. The writer believes the Grenville sub-province was the active block.

In the western part of the area the Cobalt sediments are practically flat-lying and largely undeformed. Toward the Grenville front fault zone the Cobalt series and underlying granite become progressively more deformed. The deformation does not diminish when the fault zone is crossed but continues to the east. The flat-lying Cobalt series on the other hand extends some 100 miles west of the contact of the sub-provinces indicating that this great area has been relatively free from internal deformation since Cobalt time and that it was probably the stable block.

A feature of the Temiskaming-Grenville contact is the presence of (in part erosional remnants of) late Precambrian sedimentary rocks immediately west of it: Bruce, Mistassini (same age as sediments in the Labrador Trough), Cobalt, Chibougamau, and Whitewater. These sediments may have been preserved on the west side of the zone of faulting by: (1) protection by the upfaulted block to the southeast, or (2) subsidence along parallel minor faults west of the major zone of faulting. At Lake Mistassini faults at the western ex-

tremity of the Mistassini series have the opposite relative movement to faults along the southeast margin of this series. In the area north of Killarney (Collins, 1933), most of the east-west folds in the Bruce and Cobalt series plunge eastward despite this up-faulting of the Grenville sub-province. These observations, at widely different localities, suggest a subsidence west of the main zone of faulting.

### *Lineation*

*General.*—Statistical diagrams totalling 1092 lineations were prepared. No lineation has been plotted twice. Owing to the nature of the field work, these diagrams, like those for *S*-surfaces, are not a true sample in that they do not provide an equal number of structure determinations from equal areas. However, the uniformity of the attitudes of structures has minimized this discrepancy, and certain valid conclusions can be drawn from these diagrams (Pl. 4, nos. 11–22).

Lineation, like foliation, flattens to the east. For example, maxima of lineations in the Gowganda and Lorrain formations at Rabbit Lake (Pl. 4, nos. 19, 20) plunge at 65°. To the east at Maxam Lake (Pl. 2, area 1), the plunge is 55° (Pl. 4, no. 18). For the area from Hangingstone Lake north to Tent Lake (Pl. 2, area 3) the plunge is 60°, while for the area to the east (Pl. 2, area 4) it is 40° (Pl. 4, no. 13).

The four types of $L_1$ show a remarkable correspondence: (1) long axes of mineral grains, (2) long axes of pebbles in the Gowganda formation, (3) boudinage in the Grenville subprovince, and (4) axes of minor folds in the Grenville sub-province.

*Type 1 of $L_1$ long axes of mineral grains.*—This lineation is most noticeable and best developed in the gneisses of the Grenville subprovince, especially where foliation is highly contorted or where quartz is plentiful. The quartz grains tend to be elliptical. With stronger deformation the grains become prismatic and apparently unite to form little ridges. In gneisses low in quartz lineation results from parallelism of the long axes of mica flakes. A pronounced lineation due to the parallelism of amphibole needles occurs in the amphibolites. This lineation is usually parallel to the regional lineation. However, around Red Cedar Lake it plunges southward. The amphibolites are less competent than the gneisses and probably yielded to later minor disturbances.

In the Temiskaming sub-province this lineation is found only where the rocks are affected by northeast deformation; here, and at faults in the Grenville sub-province, this lineation results from a lengthening and deforming of mineral grains parallel to $L_1$. *S*-surfaces of strongly deformed rocks high in quartz resemble the split surface of wood. Where deformation has been very intense the long axes of feldspars are often the most important contributors to lineation. In the Cobalt series at Rabbit Lake faint striations $L_1$ on bedding occur, a few plunge northwestward.

The correspondence of lineation is best in the Temiskaming sub-province and decreases with the decrease in competence of the rocks southeastward from the contact of the subprovinces. In the Grenville sub-province this lineation is imprinted throughout the gneiss. The following show the variations in a typical outcrop at Red Cedar Lake:

100°–45°E.
105°–35°E.
115°–40°E.
115°–40°E.

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PLATE 3.—PHOTOGRAPHS OF DEFORMED ROCKS

FIGURE 1.—RANKIN LAKE

Thrust faulting of granite over the younger Gowganda formation.
Beds overturned and face west. View southward.

FIGURE 2.—JUMPING CARIBOU LAKE

Thrust faulting of granite over the younger Gowganda formation.
Beds overturned and face west. View southward.

FIGURE 3.—CHARACTERISTIC TYPE OF DEFORMATION WHICH INVOLVED FLOW OF QUARTZ

From a deformed pegmatite on the west side of Temagami River 3 miles above Red Cedar Lake. Photograph by E. Elliot.

FIGURE 1

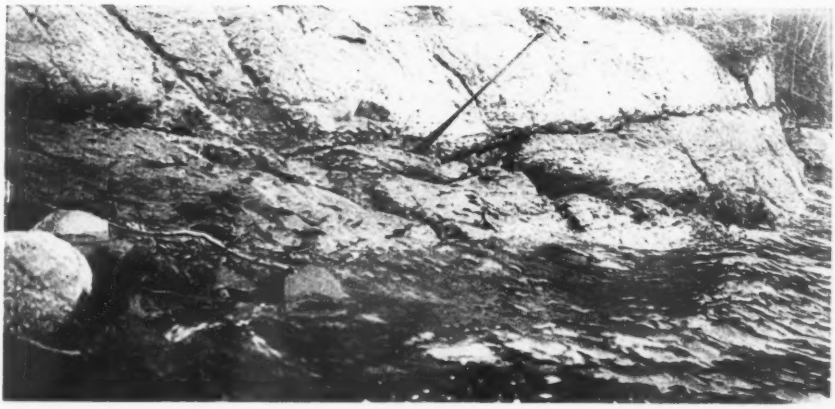

FIGURE 2

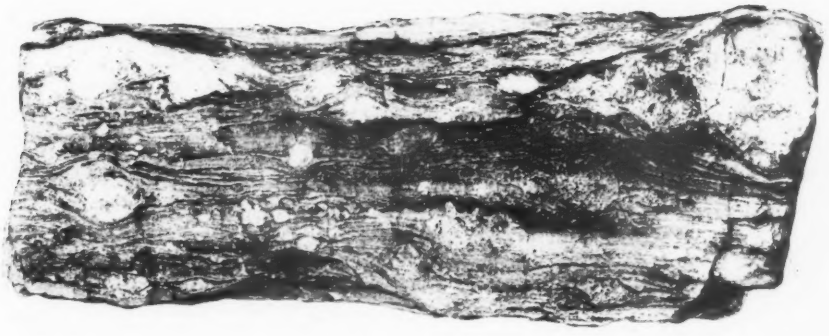


FIGURE 3

PHOTOGRAPHS OF DEFORMED ROCKS

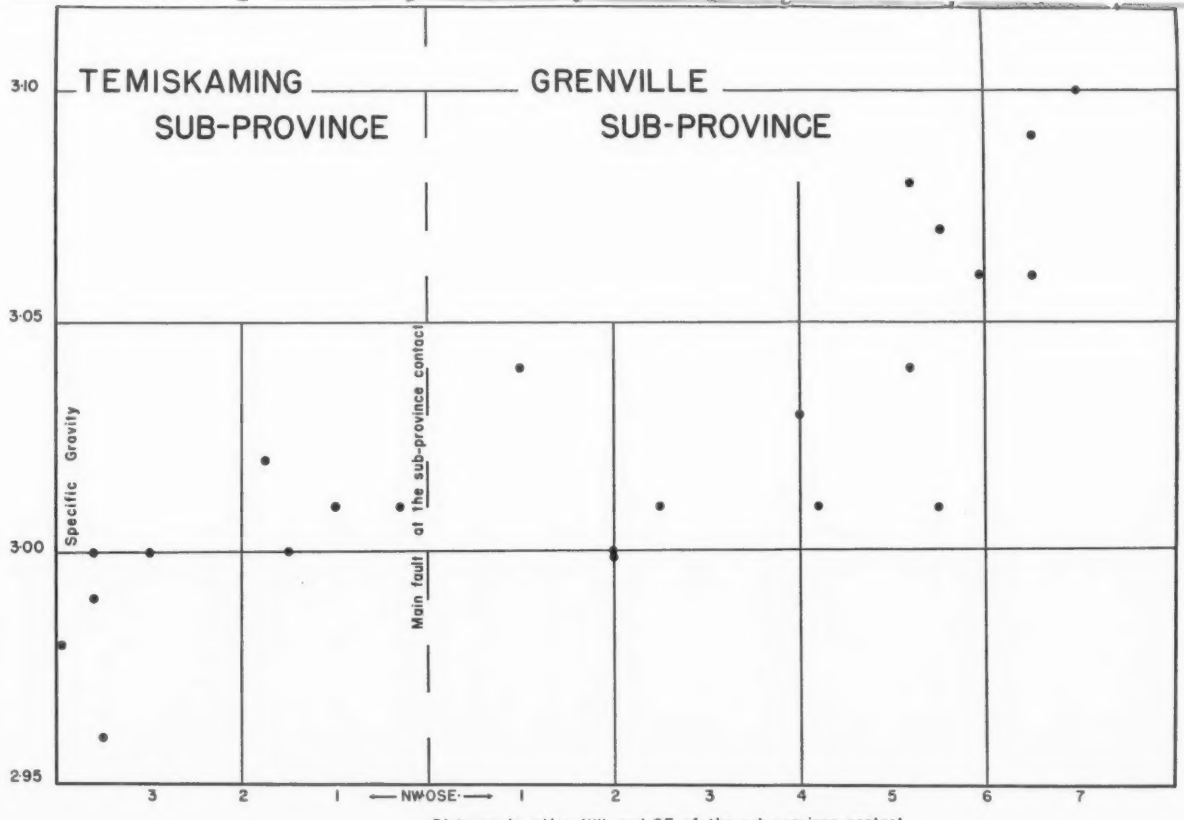


FIGURE 3.—INCREASE IN SPECIFIC GRAVITY OF OLIVINE DIABASE DIKES SOUTHEASTWARD

In any outcrop of gneiss in the Grenville sub-province the variation in azimuth of lineation is approximately twice as many degrees as the variation in plunge.

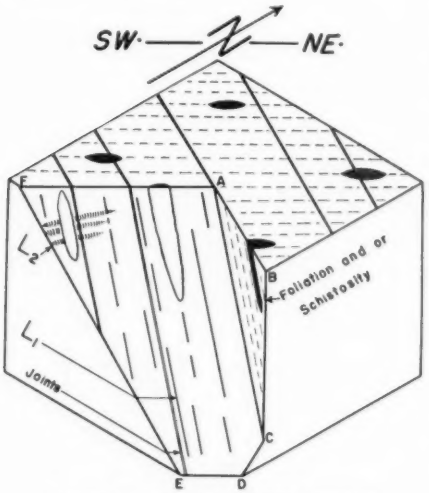


FIGURE 4—ISOMETRIC BLOCK DIAGRAM

Showing relation of structural features. Plane ACDEF is the plane of northeast-southwest schistosity. Plane ABC represents prominent joints at right angles to schistosity; AC, the line of intersection of *S*-surfaces, is parallel to $L_1$, represented by mineral elongation and long axes of pebbles, boudinage, and axes of minor folds.

*Type 2 of $L_1$ long axes of pebbles in Gowganda Formation.*—This lineation (Pl. 4, no. 16) occurs in the Gowganda conglomerate from Rankin Lake (Pl. 2, area 2) northeastward in what is the farthest eastward projection of these rocks from the normal northeast trend of their eastern limit. This is also the most highly deformed area of these rocks.

The pebbles (average estimated length 2 inches) appear to have migrated to a position with long axes parallel to $L_1$. In a few cases, such as at Rankin Lake, where deformation has been very intense, the pebbles have been elongated parallel to $L_1$ and $L_2$.

*Type 3 of $L_1$ boudinage in Grenville sub-province.*—Only 45 occurrences of boudinage were recorded for the area (Pl. 4, no. 15). The boudins are high in amphibole and have the same composition as the amphibolites of the area, except for a few that have an even higher percentage of amphibole. The boudins are flattened cylindroids with their long axes parallel to $L_1$ and their intermediate axes parallel to $L_2$. They range considerably in size; the larger ones are about 4 inches in a direction parallel to $L_1$ and 1½ inches in a direction normal to the plane $L_1L_2$ (Fig. 4). Some boudins were undoubtedly basic sills parallel to the foliation and later stretched and separated into these forms. A large boudin of porphyritic diabase with minor dimensions 4 feet by 1 foot occurs on the east side of the highway, half a mile north of Marian Lake (Pl. 2, area 4). Some boudins may have been undigested inclusions in granite.

*Type 4 of $L_1$ axes of minor folds in Grenville sub-province.*—This lineation is always accompanied by and parallels lineation formed by long axes of mineral grains (Pl. 4, nos. 11, 12). It occurs only in the rocks of the Grenville sub-province. These folds are usually *S*-shaped and have an amplitude from a fraction of an inch to over 3 feet. They are not confined to any particular type of gneiss.

### *Lineation $L_2$ Minute Ridges and Wrinkling Axes at Right Angles to $L_1$*

The best exposures of this, the least common lineation, occurs in rock cuts in the granite where large surface areas along northeast schistosity are exposed. $L_2$ in the granite forms minute discontinuous ridges with horizontal or nearly horizontal axes. The lower sides of the ridges are more strongly marked by lineation $L_1$ than the upper sides. There has been an appreciable flowage of the quartz in the granites in which it occurs. The ridges seemingly were formed by a migration of the feldspars, moving parallel to $L_1$ and coming to rest in the horizontally protected areas on either side of a larger crystal of feldspar. Feldspar was the most resistant mineral to deformation. Analogous structures were observed in the Gowganda conglomerate where horizontal wrinkling of schist lamellae occurs in the protected areas on either side of pebbles. However, $L_2$ in the Gowganda formation is most commonly a wrinkling of schist lamellae. The feldspars and pebbles do not appear to have rotated during the formation of $L_2$, nor does rotation seem to have played a part in the deformation.

*Analysis of Lineation*

In addition to the correspondence in attitude of lineation and foliation in both sub-provinces, two other structural relationships hold rigidly throughout the area: (1) lines of intersection of the northeast schistosity and the northwest joints are parallel to $L_1$ (*compare* Pl. 4, nos. 23, 24, 25 lines of intersection of *S*-surfaces on individual outcrops *with* Plate 4, nos. 11–22 for lineation $L_1$); (2) pebbles in the deformed Gowganda conglomerate, quartz grains, boudinage, and feldspar augen in the Marten River granite all have their long axes parallel to $L_1$, their intermediate axes parallel to $L_2$ and their minor axes at right angles to the northeast-striking eastward-dipping *S*-surfaces. This shortening axis at right angles to the northeast *S*-surface (plane $L_1$ $L_2$) is in accordance with a tension origin for the northwest joints.

The evidence, discussed in the section on the Direction of Tectonic Transport, indicates movement parallel to $L_1$. The only feature not at once reconcilable with this is the folding in the gneisses along axes parallel to $L_1$ (Pl. 4, nos. 11, 14). Although these minor folds in the gneisses may have originated earlier than other types of $L_1$ and owe their origin to another cause, it seems best to explain them by the same movement, because all types of $L_1$ are constant in attitude. Two conceivable ways these folds could have formed are: (1) more rapid movement or more movement of closely adjacent parts of the same "bed" (couple action) with continued movement parallel to $L_1$; folds thus formed would tend to migrate to a position parallel to $L_1$; (2) lengthening of the "beds" parallel to the strike line ($L_2$) due to flattening as seen in quartz grains and boudins; this lengthening of beds in the direction of $L_2$ could have been accommodated by folding adjacent to the areas undergoing flattening. A convergence of lineation $L_1$ was not observed, and this condition as described by Cloos (1946, p. 28) does not appear to have been the cause of the folds in this case.

It might be argued that these folds ($L_1$) occur around Red Cedar Lake because $L_1$ does not coincide with the dip line of the foliation, as it does in the rest of the area, but plunges east of it. This possibility has been investigated by a method proposed by Sander, described by Fairbairn (1949, p. 194). Each *s*-surface is plotted (with its lineation $L_1$ long axes of mineral grains) and then rotated on a vertical axis until the *s*-surface strikes 90° and dips south. The position of the lineation for the rotated position of its *s*-surface is transferred to a new diagram. These diagrams (Pl. 4, nos. 26–29) indicate that the divergence of the plunge of lineation, east of the dip line of the foliation or schistosity, was not the cause or even a factor in the formation of these folds. They form where the dip line is both parallel to the lineation (Pl. 4, nos. 14, 27) and oblique to it (Pl. 4, nos. 11, 28).

This discordance of the lineation with the dip line in the Red Cedar Lake area, in contrast to the parallelism for the remainder of the area, suggests deformation of a slightly different attitude from that in the rest of the area, or a tilting of this area after the formation of $L_1$.

Brief examination suggests that the same structures as those in the map area occur along the contact of the sub-provinces at least as far as Sudbury. Observed lineations are shown on Plate 1. The few lineations (Pl. 1) seem to indicate that in the region some distance southeast of the contact of the sub-provinces movement was westward rather than northwest. The writer interprets the west movement as earlier than the northwest movement that produced faults and caused post recrystallization deformation along the western limit of the Grenville sub-province.

Haig township (Quebec, Pl. 1) is the only other area at the contact of the sub-provinces for which data on lineation is at present available (Norman and Dawson, 1947). It is difficult to interpret their observations because the lineation as mapped transects the plan of northeast schistosity.

Wahl and Osborne (1950) mapped an area within the Grenville sub-province about 10 miles southeast of the Temiskaming-Grenville contact (Pl. 1); however, they did not offer an interpretation of the structural features. The one given for the Maxam Lake Cross-Lake area would seem to fit their observations. A statistical diagram of the lineations (similar to lineation $L_1$ of the writer) shown on their map was compiled by the writer (Pl. 4, no. 30). The plunge is much flatter than in the area studied

by the writer, agreeing with the conclusion that lineation flattens toward the east. In the area mapped by Wahl and Osborne, westward-plunging lineation is shown only on the east side of an inverted "cone" structure. It is also noticeable from their map that along certain zones, marked by northward-trending elongate lakes perhaps lying along faults, the lineation has a much steeper plunge than the regional lineation. These probable faults strike northerly, and the lineation strikes east indicating movement from east to west rather than predominantly from southeast to northwest as in the area described by the writer. Joints parallel to lineation $L_1$ are also common in this area, and boudinage occurs (Wahl and Osborne, 1950, Pl. iv).

### Igneous Structures

There is great difficulty in distinguishing primary from secondary structures in granite of the Grenville sub-province. All northeast structures have the same attitude, and igneous and secondary structures originated from movement in the same direction.

The Marten River granite appears to be decidedly separate from granite 2. Its strong lineation, resulting from a parallel orientation of large feldspar augen, is apparently primary. Its correspondence with regional lineation and its northeast trend suggest a syntectonic origin perhaps accompanying one of the later deformations.

### Age of Structures

Movement on northeast structures has taken place at least as many times as there are rocks of different ages to date it.

In areas of granite 1, granite 2 has been emplaced along northeast fractures. Major northeast faulting preceded deposition of the Cobalt series (Todd, 1925, p. 9). In the largest bay on the south side of Southeast Bay, Rabbit Lake, undisturbed Gowganda formation overlies a highly sheared granite. At Rankin Lake the Gowganda formation occurs on the east side of a major fault (Pl. 2). If the displacement (east side up) is as great as indicated by deformation in the granite, it is difficult to see how this area of Gowganda formation could have escaped erosion, unless deposition took place in a pre-Cobalt valley along a fault which subsequently became active. A small island of conglomerate in Rankin Lake has granite boulders that were sheared before deposition because the attitude of schistosity varies greatly from one boulder to the next.

At Jumping Caribou Lake (Pl. 2, W. of area 3) a great deal of soft-rock deformation in the Gowganda formation, near the fault at the southeast part of this lake, suggests movement on these faults after deposition and prior to consolidation of this formation.

The difference in character of the sediment forming the Gowganda and Lorrain formations suggests faulting in the Grenville sub-province between the deposition of the Gowganda and the Lorrain formations.

The correspondence between attitudes of foliation, schistosity, joints, and lineation in both sub-provinces (Pl. 4, nos. 1–29) shows that the same late deformation affected both the gneisses of the Grenville sub-province and the Cobalt series. In rocks older than the Cobalt series this deformation followed earlier structures.

The Nipissing diabase shows undisturbed chilled contacts against heavily sheared granite along the lumber road east of Rabbit Lake and along number 11 highway between Jumping Caribou and Olive Lakes. Plate 4 (no. 10) shows poles to *S*-surfaces in the Nipissing diabase, which differ from the diagrams for other rocks. Although this is conceivably due to differences in physical properties between the diabase and other rocks, it more probably indicates that the Nipissing diabase in the Temiskaming sub-province has not been subjected to all the deformations that affected the older Cobalt series. However, in the Grenville sub-province where the Nipissing diabase was not identified as such (probably because it has been converted to amphibolite), the younger olivine diabase has been deformed and in places converted to amphibolite indicating post-olivine diabase deformation in the Grenville sub-province.

Where olivine-diabase dikes are absent in the Grenville sub-province, the deformation can be separated into pre- and post-recrystallization (metamorphism). The post-recrystal-

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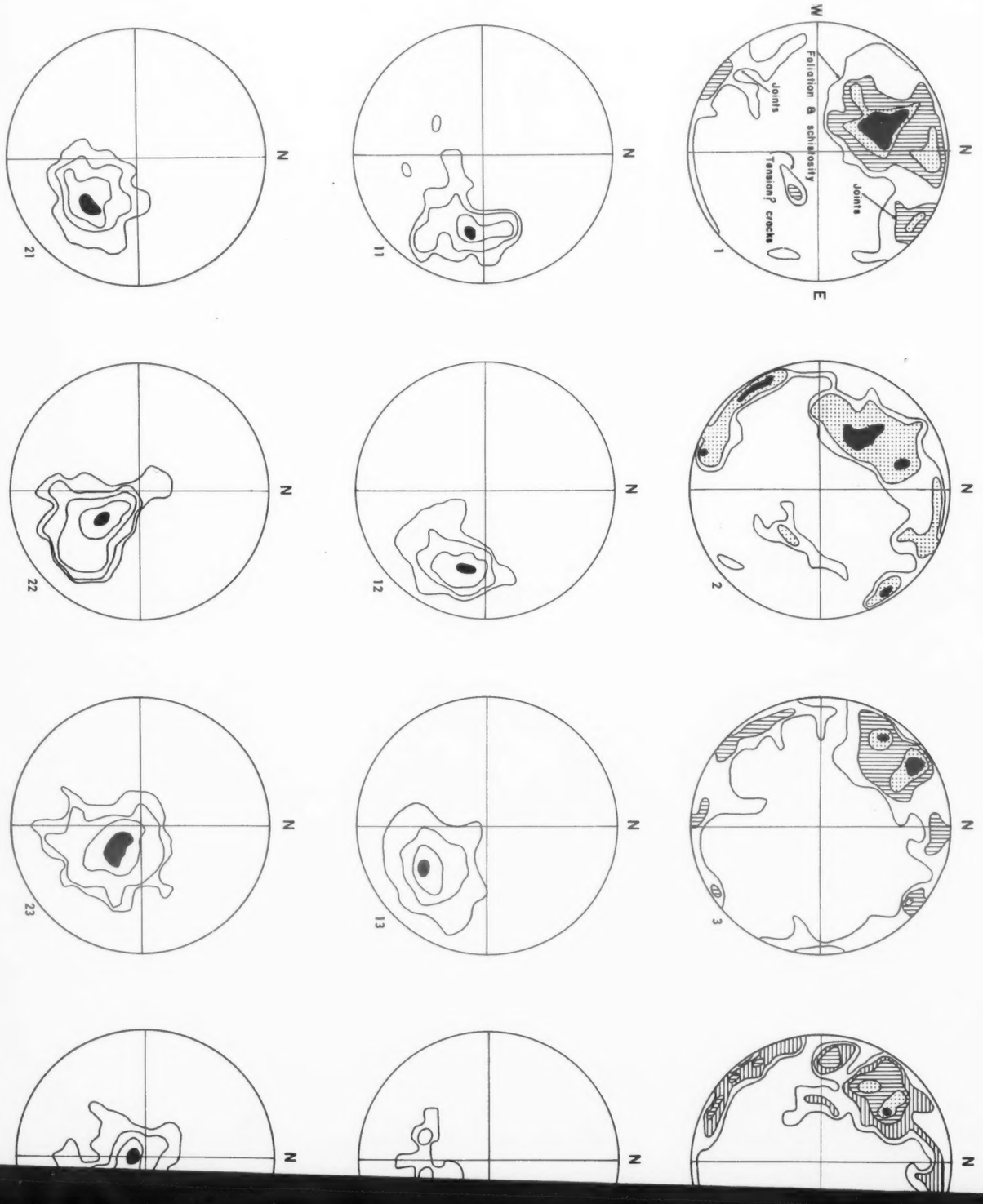
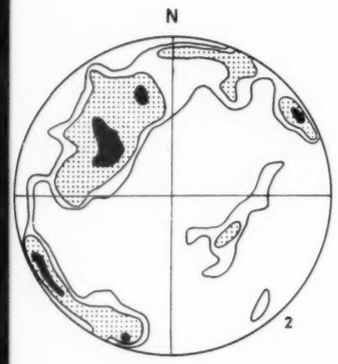
| No. | Structure | Area No. Plate 2 | Area | Formations | No. of Readings | Contours % |
|---|---|---|---|---|---|---|
| 1 | poles to *s*-surfaces | 6 | Red Cedar Lake to Marian Lake | granite and gneiss | 325 | 4, 3, 2, and 1 |
| 2 | poles to *s*-surfaces | 5 | Cross Lake and vicinity | all formations | 230 | 4, 2, and 1 |
| 3 | poles to *s*-surfaces | 3 | Hangingstone Lake to Tent Lake | granite and gneiss | 275 | 6, 4, 2, and 1 |
| 4 | poles to *s*-surfaces | | South of the railroad except at Rankin Lake | Gowganda | 110 | 9, 6, 4, 2, and 1 |
| 5 | poles to *s*-surfaces | 4 | Opechee Lake north to the railroad | granite and gneiss | 372 | 7, 5, 3, 2, and 1 |
| 6 | poles to *s*-surfaces | | Rabbit Lake | Lorrain | 130 | 11, 7, 5, 3, and 1 |
| 7 | poles to *s*-surfaces | | North of the railroad and at Rankin Lake | Gowganda | 300 | 5, 4, 2, and 1 |
| 8 | poles to *s*-surfaces | 2 | Rankin Lake north to the lumber road | granite | 190 | 7, 6, 4, 2, and 1 |
| 9 | poles to *s*-surfaces | 1 | Maxam Lake and vicinity | all formations | 140 | 5, 3, 2, and 1 |
| 10 | poles to *s*-surfaces | | Entire area | Nipissing diabase | 160 | 4, 3, 2, and 1 |
| 11 | $L_1$ axes of minor folds | 6 | Red Cedar Lake to Marian Lake | granite and gneiss | 90 | 20, 10, 5, and 2 |
| 12 | $L_1$ long axes of mineral grains | 6 | Red Cedar Lake to Marian Lake | granite and gneiss | 240 | 20, 10, 5, and 2 |
| 13 | $L_1$ long axes of mineral grains | 4 | Opechee Lake north to the railroad | granite and gneiss | 225 | 15, 10, 5, and 1 |
| 14 | $L_1$ axes of minor folds | 4 | Opechee Lake north to the railroad | granite and gneiss | 45 | 20, 15, 5, and 2 |
| 15 | $L_1$ boudinage | | Grenville sub-province part of the map area | gneiss | 45 | 20, 10, 5, and 2 |
| 16 | $L_1$ long axes of pebbles | | Rankin Lake and north of the railroad | Gowganda | 52 | 30, 20, 10, and 5 |
| 17 | $L_1$ long axes of mineral grains | 2 | Rankin Lake north to the lumber road | granite | 72 | 25, 15, 5, and 3 |
| 18 | $L_1$ long axes of mineral grains | 1 | Maxam Lake and vicinity | all formations | 38 | 20, 15, 10, 5, and 3 |
| 19 | $L_1$ long axes of mineral grains | | Rankin Lake | Lorrain | 37 | 20, 10, 5, and 3 |
| 20 | $L_1$ long axes of mineral grains | | Rankin Lake and north of the railroad | Gowganda | 66 | 20, 15, 10, 5, and 2 |
| 21 | $L_1$ long axes of mineral grains | 5 | Cross Lake and vicinity | all formations | 90 | 20, 10, 5, and 1 |
| 22 | $L_1$ long axes of mineral grains | 3 | Hangingstone Lake to Tent Lake | granite | 92 | 15, 10, 5, 2, and 1 |
| 23 | lines of intersection of *s*-surfaces | | Entire area | Gowganda and Lorrain | 219 | 10, 5, 2, and 1 |
| 24 | lines of intersection of *s*-surfaces | 5, 3, 2 and 1 | Cross Lake, Hangingstone Lake to Tent Lake, Rankin Lake north to the lumber road, and Maxam Lake | all formations | 227 | 7, 5, 3, and 1 |
| 25 | lines of intersection of *s*-surfaces | 6 and 4 | Red Cedar Lake and Opechee Lake areas | granite and gneiss | 164 | 5, 3, 2, and 1 |
| 26 | Position of $L_1$ after rotating *s*-surfaces to east-west strike and south dip | | Entire area | Gowganda and Lorrain | 130 | 20, 10, 5, and 1 |
| 27 | same as 26 | 5, 3, 2 and 1 | Cross Lake, Hangingstone Lake to Tent Lake, Rankin Lake, north to the lumber road, and Maxam Lake | granite and gneiss | 196 | 18, 10, 5, and 1 |
| 28 | same as 26 | | Red Cedar Lake to Marian Lake | granite and gneiss | 188 | 10, 8, 5, 2, and 1 |
| 29 | same as 26 | | Opechee Lake north to the railroad | granite and gneiss | 148 | 15, 10, 5, and 2 |
| 30 | "L"$_1$ Q.D.M. Map 820 | | Cawatose Area | granite and gneiss | 255 | 12, 8, 6, 4, 2, and 1 |

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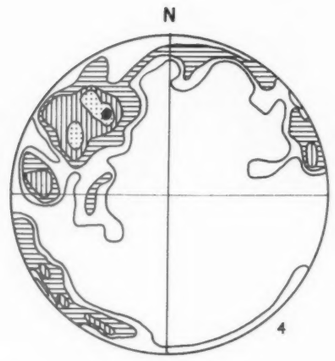
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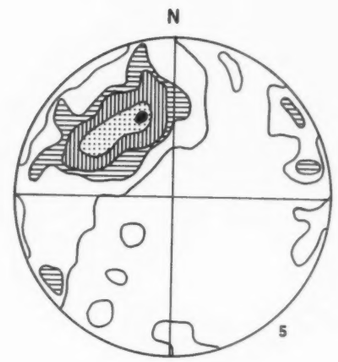
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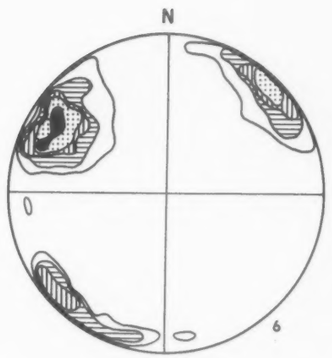
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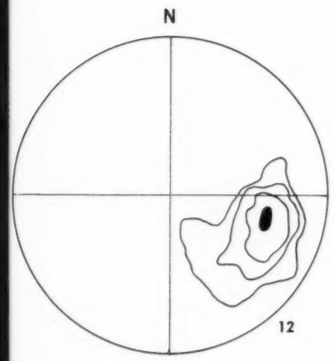
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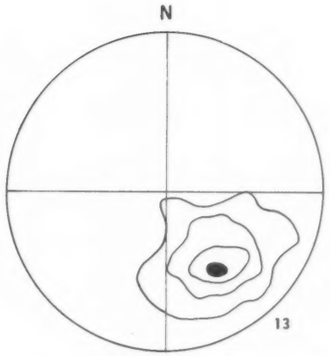
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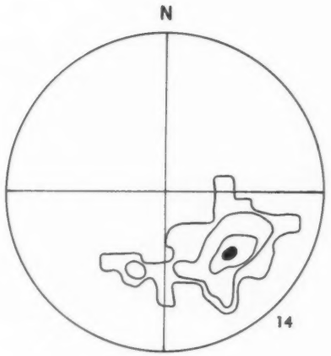
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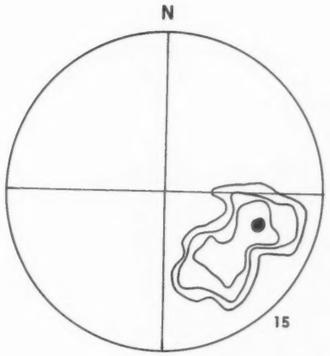
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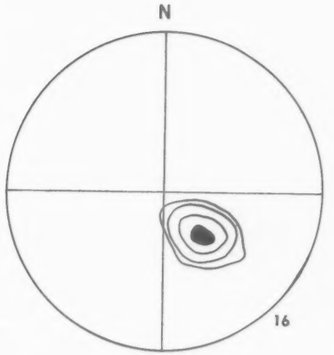
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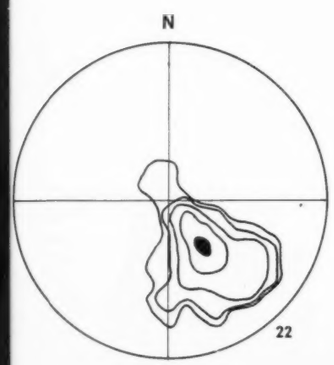
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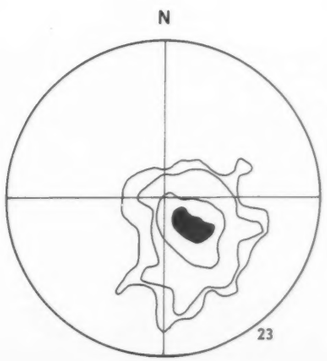
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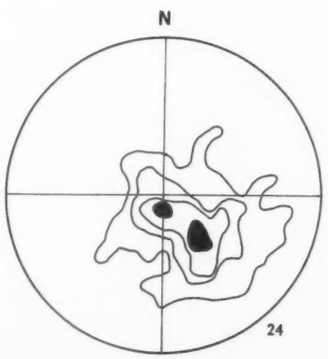
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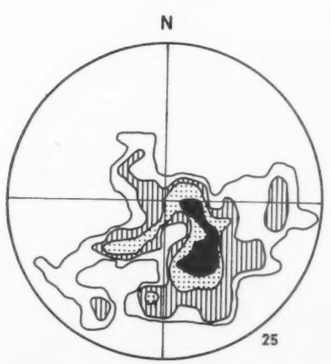
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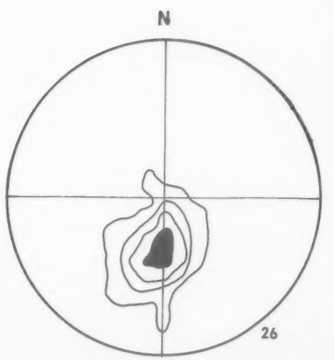
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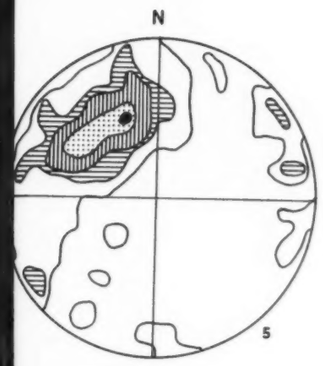


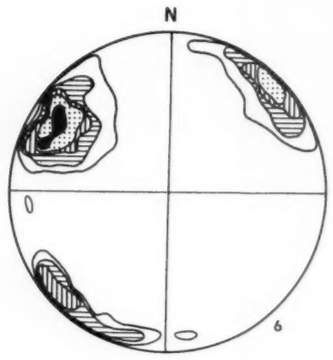


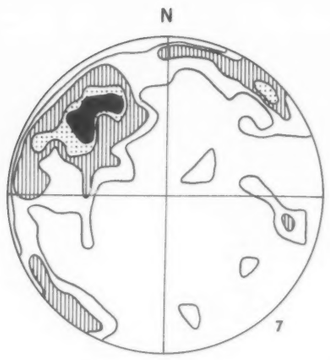


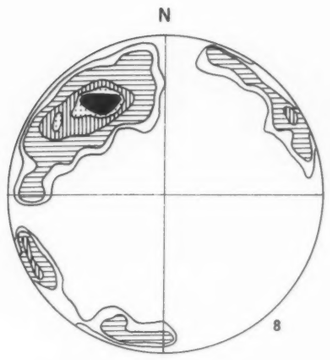


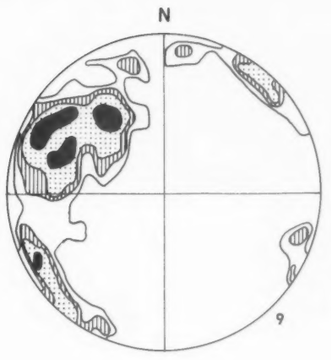


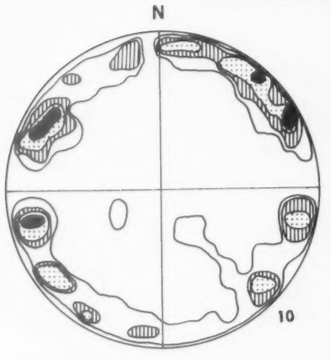


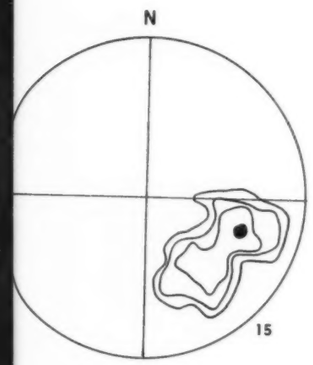


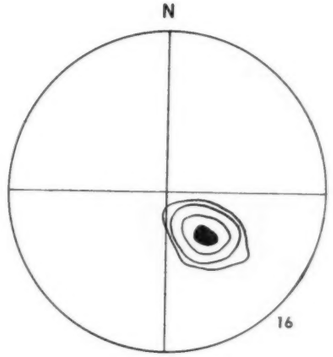


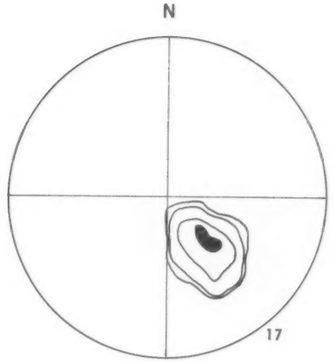


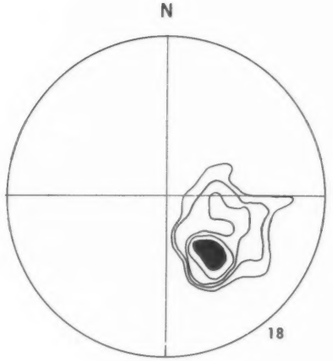


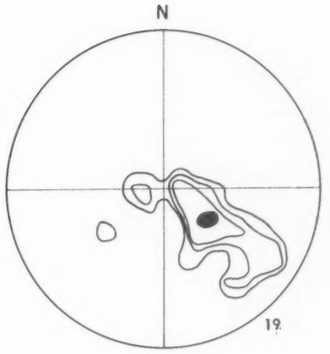


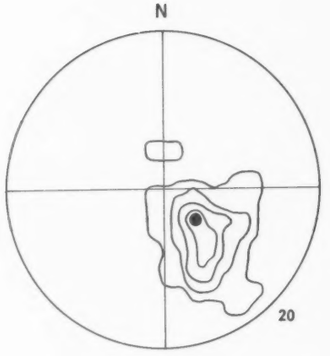


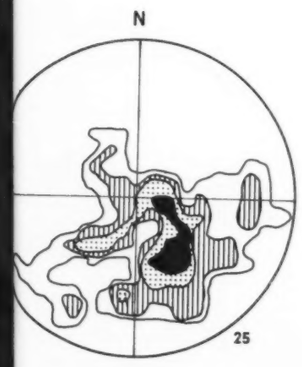


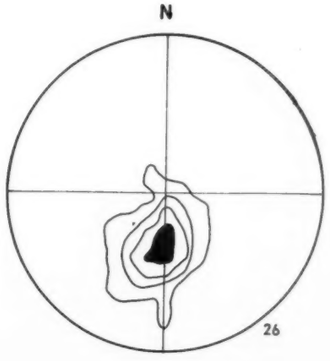


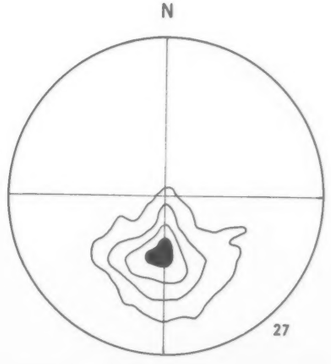


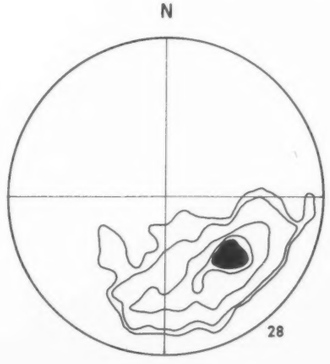


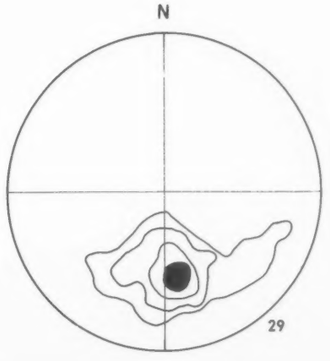


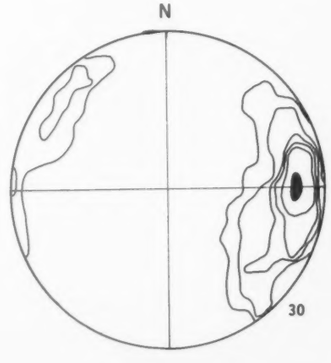


STATISTICAL DIAGRAMS OF STRUCTURES

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lization deformation has produced a discernible cataclastic texture and schistosity.

Northwest structures predate the Nipissing diabase and have in part controlled its intrusion and, more markedly, intrusion of the olivine diabase. North of the area, around Lake Temiskaming, post-Silurian movement has taken place on faults trending northwest. According to Hume (1925, p. 51) this movement (east side down about 1000 feet) possibly took place in the Pliocene. Movement is still taking place on the Ottawa river fault (Hodgson, 1935, p. 438; 1937, p. 116). At Red Cedar Lake north-south fractures are offset by movement along northwest fractures.

Although faults, schistosity, and joints of all three directions of strike (Figs. 1, 4) have had a long and complex history, the northwest-trending faults (*e.g.*, Ottawa River fault) are later than the northeast structures, for they followed joints which formed parallel to movement on the northeast faults.

## Structural Petrology

Quartz, epidote, and feldspar are the only minerals present in sufficient quantity for optical fabric study. Biotite is uniaxial and has a very strong dimensional orientation parallel to the foliation (diagrams not reproduced here). Fabric diagrams were not prepared for plagioclase, but it appears to have a preferred orientation in the Grenville-type gneisses. Of 47 bisectrix figures obtained on plagioclase in oriented sections of Grenville-type gneiss, 40 had *Z* parallel with $L_1$, and only 7 had *X* parallel with $L_1$. *Y* tends to be parallel to $L_2$. The average composition of the plagioclase in these thin sections is $An_{26}$.

Two oriented sections contained enough quartz and epidote to make fabric diagrams for both (Pl. 5, nos. 37, 47, 50, 51). *Y* of the epidote is parallel to $L_1$, and these diagrams for epidote resemble those obtained by Ladurner (1951, p. 334) for this mineral.

Fabric diagrams for quartz (Pl. 5, nos. 31–49) are all contoured at 1 per cent intervals; the lowest contour is 1 per cent. These diagrams are northeast-southwest, more or less complete girdles, with the lineation $L_1$ (direction of movement) as axis. Hence they are similar to diagrams obtained from somewhat similar structures by Kvale (1941, p. 191), Strand (1944, p. 24), Cloos (1947, p. 6), and Balk (1952, p. 421). Seemingly quartz girdles from the Moine schists of Scotland, have the same symmetry and attitude with respect to lineation as those in the present area and are also at right angles to movement rather than parallel as maintained by Phillips (1937, p. 591; 1945, p. 207).

The quartz fabric is symmetrical about $L_1$. Although many refuse to accept quartz girdles at right angles to movement it seems highly improbable that a late movement of a few mm perpendicular to $L_1$ could have completely reoriented the quartz. If quartz can be reoriented by movement so slight that no megascopic evidence results, the study of quartz fabric is useless.

In the quartz diagrams there tends to be a break or a lowering of concentration at the schistosity and on each side of the girdle, at right angles to the foliation or schistosity. The angle between the foliation and the maxima or highs nearest the foliation is near 38°13′—the angle between the *m* and *r* faces of quartz. Similarly the larger angle between the next concentrations and the foliation averages near 66°52′—the angle between the *m* and *z* faces of quartz. This suggests that the *m* and *z* planes may have played a part in the orientation of the quartz. Fairbairn (1939, p. 363) has pointed out that *r* and *z* should be the best "cleavages" in quartz from a structural basis. Percentage of quartz relative to other minerals seems to have exerted no control on the quartz orientation. There is a strong similarity among all diagrams, and hence diagrams for rocks of different age, origin, composition, and periods of deformation are similar.

Quartz deformation lamellae occur in rocks with cataclastic texture and are especially common in the Cobalt series, possibly because intergranular movement could take place more easily in these rocks owing to the noninterlocking of the grains and weaker cementing between mineral grains than in the granites and gneisses. There were sufficient lamellae in one specimen to prepare a diagram of their poles (Pl. 5, no. 49).

In the Temiskaming sub-province only sheared rocks show preferred orientation of

quartz. In the Grenville sub-province all rocks have preferred orientation of quartz whether they have a cataclastic texture or not. Hence the quartz fabric is the same for pre- and post-recrystallization movement. The pre-recrystallization movement in the Grenville sub-province took place under conditions different from the movement in the Cobalt series, as can be seen from the types of structures developed in the Grenville sub-province, such as boudinage and minor folds with axes parallel with $L_1$. However, these entirely different structural and metamorphic types have the same quartz fabric. Does this mean the quartz fabric is independent of depth of burial and metamorphism, and is dependent entirely on tectonic history?

## Metamorphism

### *General Statement*

In granite 1 in the Temiskaming sub-province the "saussuritization" of the plagioclase and alteration of the hornblende to chlorite and epidote increased with the degree of deformation. The more calcic the plagioclase the more susceptible it was to "saussuritization". In "saussuritized" plagioclase the zoisite is more abundant at the center of the plagioclase, and white mica at the outside. Biotite was much less susceptible to alteration to chlorite than hornblende to chlorite and epidote. These alterations seem to be of a regional nature and are probably characteristic of the Temiskaming sub-province as a whole.

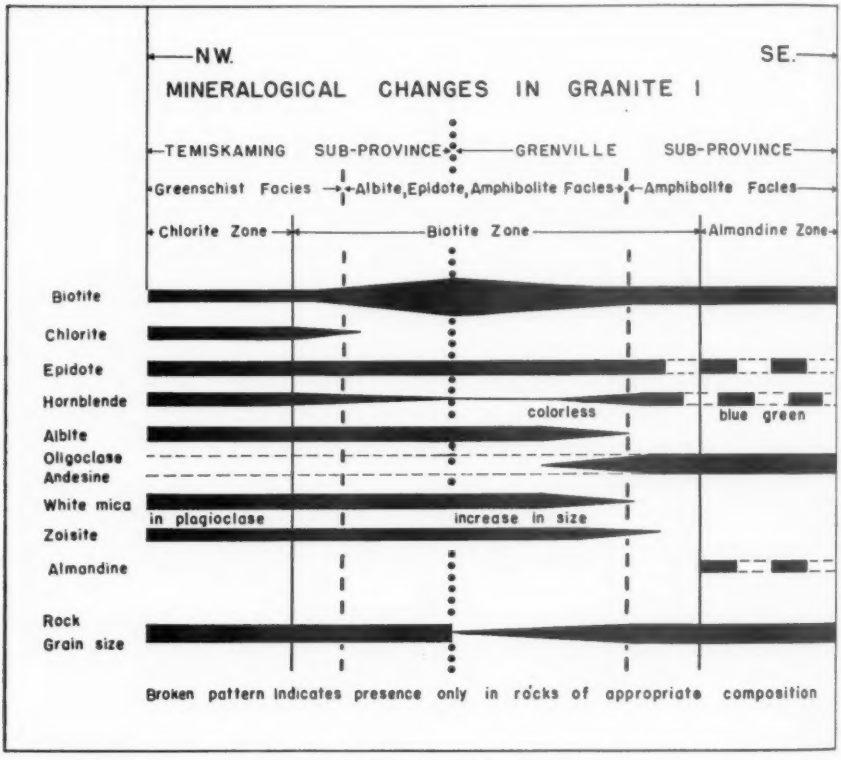


FIGURE 5.—METAMORPHISM OF GRANITE 1

The location of the biotite isograd and the various mineral assemblages are shown on Plate 2. Figure 5 depicts mineralogical and other changes from northwest to southeast. The position of the biotite isograd in the Temiskaming sub-province is related to the contact of the sub-provinces, and this contact can be determined most accurately by the degree of recrystallization.

There are certain difficulties in the study of metamorphism in granite 1: (1) It did not reach equilibrium in the chlorite zone (greenschist facies) before metamorphism. The plagioclase with few exceptions did not completely alter to $An_0$. (2) Rocks in the amphibolite facies are probably the only ones that reached equilibrium during metamorphism. (3) The coarse grain of these rocks has made them less sensitive to metamorphism. (4) The faulting, which is later than metamorphism, has in some cases brought areas of different grades of metamorphism into juxtaposition. And (5) later granitization has left unaffected few complete sections parallel to the increase in grade of metamorphism.

The biotite isograde in granite 1 was located by study of thin sections. In the biotite zone secondary biotite is in fine aggregates. Relicts of unaltered hornblende, in the same optical orientation, are surrounded by biotite containing epidote. This biotite formed from the chlorite produced by the previous alteration of the hornblende. In the biotite zone recrystallization of chlorite to biotite, and of the epidote-zoisite-white mica alteration of the plagioclase to coarser and fewer grains took place. The hornblende not previously altered appears stable in this zone; where there has been intense shearing hornblende is absent although epidote remains. To the east, in what appears to be the albite-epidote amphibolite facies, the hornblende is surrounded by a colorless amphibole, and both types have the same optical properties and orientation.

If data were available to locate the garnet isograd it would probably parallel the biotite isograd and the contact of the sub-provinces. Faulting has disrupted the garnet isograd more than the biotite isograd. For the above reasons the boundaries between the albite-epidote amphibolite facies and amphibolite facies could not be accurately determined; hence individual locations of the facies and assemblages shown.

TABLE 1.—MINERAL ASSEMBLAGES IN AMPHIBOLITE FACIES

| | No. of plagioclase determinations | Average An Content |
|---|---|---|
| oligoclase-biotite-quartz | 6 | 25 |
| oligoclase-biotite-quartz-epidote | 7 | 26 |
| oligoclase - biotite - quartz - epidote-hornblende | 7 | 26 |
| andesine - biotite - quartz - hornblende | 5 | 30 |
| andesine - biotite - quartz - almandine | 6 | 32 |
| oligoclase - biotite - hornblende-almandine ± quartz | 2 | 28 |

### *Amphibolite Facies*

Granite 1 (Pl. 2, pattern 4a) in the Grenville sub-province is in the amphibolite facies. In this facies the white mica has gone into the biotite, and the zoisite has with few exceptions gone back into the feldspar.

The following evidence suggests that the gneisses in the Grenville sub-province portion of the area are in part at least metamorphosed granite 1: (1) gneisses of the Grenville sub-province and granite 1 are similar in chemical composition (Fig. 2); and (2) granite 1 can be traced from the Temiskaming sub-province through increasing degrees of metamorphism to the amphibolite facies. Through this change there is, admittedly, an abrupt increase in the degree of recrystallization as one crosses faults going southeastward. However, relict textures, such as alteration products in the feldspars, in various degrees of recrystallization can be seen on either side of the contact of the sub-provinces.

In studying the amphibolite facies specimens were selected that showed no granitization or post-recrystallization deformation. It is thought the assemblages determined were free from granitization except for the introduction of a

minor amount of material to form apatite, sphene, pyrite, and occasional carbonate. Assemblages in probable order of abundance are listed in Table 1.

A metamorphosed diabase, specimen 47-86 from number 11 highway at the east limit of the map area (Pl. 2) had the assemblage andesine-hornblende-scapolite-quartz. An assemblage containing both almandine and epidote was not found.

The abundance of the simple mineral assemblage oligoclase-biotite-quartz and the same mineral assemblages in rocks of varying composition suggest these rocks reached equilibrium.

Table 2.—Indices of Biotites

| Specimen No. | Xd | Yd | Bi | Rock type or mineral assemblage |
|---|---|---|---|---|
| 48-413 | 1.599 | 1.651 | .052 | granitized gneiss |
| 49-103 | 1.595 | 1.647 | .052 | andesine-quartz-biotite-almandine |
| 49-210 | 1.582 | 1.631 | .049 | andesine-quartz-biotite-almandine |
| 49-188 | 1.581 | 1.631 | .050 | oligoclase-biotite-hornblende-almandine-quartz |
| 47-81 | 1.585 | 1.635 | .050 | oligoclase-quartz-biotite-epidote |
| 48-364 | 1.583 | 1.630 | .047 | oligoclase-quartz-biotite |
| 48-356 | 1.581 | 1.631 | .050 | oligoclase-quartz-biotite |
| 48-202 | 1.579 | 1.624 | .045 | andesine-quartz-biotite-hornblende |
| 48-156 | 1.580 | 1.628 | .048 | albite-epidote amphibolite-facies |
| 49-253 | 1.585 | 1.641 | .056 | oligoclase-biotite-epidote-quartz |
| 48-309 | 1.593 | 1.647 | .054 | amphibolitized olivine diabase |
| 49-208 | 1.591 | 1.639 | .048 | granite 2 |

### *Minerals in Amphibolite Facies*

Biotite occurs in all rocks except the assemblage containing scapolite. It occurs in varying amounts up to 15 per cent—usually between 5 and 10 per cent. The small amount of $K_2O$ in these rocks is in the biotite. Neither microcline nor orthoclase was found in thin sections of the amphibolite facies. Dimensions of biotite flakes are in the ratio 10:3:1 in the directions parallel with $L_1$, parallel with $L_2$, and at right angles to the plane $L_1$ $L_2$ respectively. It is uniaxial. The color of the biotite is fairly uniform; *X* is pale yellowish brown, and *Y* is deep brownish red. Pleochroic halves in the biotite are common, especially near zones of faulting. They are up to 0.04 mm in diameter and are caused by minute, usually equidimensional, sphene grains or occasionally, allanite.

No obvious correlations of biotite indices and rock type or grade of metamorphism is apparent from the data of Table 2.

Epidote occurs in well-defined grains, usually with the same interference color throughout. It is almost always associated with biotite. The widespread occurrence of epidote suggests it must be part of a stable assemblage in the amphibolite facies (Turner, 1948, p. 88).

Hornblende seldom occurs in amounts of more than 10 per cent except in amphibolites and amphibolite rock forming boudins. The average composition of the plagioclase from seven specimens of gneiss containing hornblende was $An_{24}$. There is probably considerable variation in the composition of the amphibole as the extinction angle $Z \wedge C$ varies from 13° to 22°. Pleochroism is *X* yellow, *Y* emerald green, and *Z* dark green or bluish green. Hornblendes with a smaller extinction angle have pleochroism in bluish greens. Some have pleochroic haloes caused by small sphenes.

Plagioclase occurs in all rocks of the area. Grain shapes have the approximate dimension ratios 3:1:1 for dimensions parallel to $L_1$, parallel to $L_2$, and perpendicular to the plane $L_1$ $L_2$. Albite twinning is slightly more common than pericline twinning in the 15–35 per cent of the grains that are twinned. Porphyroblasts of plagioclase up to 5 mm in diameter are found in rocks with cataclastic texture, while

they are much smaller elsewhere, suggesting that deformation aided their growth.

Almandine occurs in amounts less than 5 per cent except for a few amphibolites having up to 10 per cent almandine. It was found in the rocks with plagioclase of highest average An content ($An_{31}$). Almandines are equidimensional and average about 1 mm in diameter. In the gneisses they occur as trapezohedrons with striations on the faces. They often have inclusions of quartz and feldspars. None examined showed snowballing. It seems remarkable that the same (?) garnet almandine formed in rocks of such different age and composition as seen from Table 3.

TABLE 3.—PROPERTIES OF GARNETS

| Specimen no. | Index | S.G. | Unit cell Å | Mineral assemblage or rock |
|---|---|---|---|---|
| 49-210 | 1.79 | 4.04 | 11.57 | andesine-biotite-quartz-almandine |
| 48-305 | 1.80 | 4.13 | 11.53 | andesine-biotite-quartz-almandine |
| 47-86 | 1.80 | 4.10 | | andesine-biotite-quartz-almandine |
| 49-188 | 1.79 | 4.20 | 11.63 | andesine-biotite-quartz-almandine |
| 49-185 | 1.80 | 4.16 | 11.56 | andesine-biotite-hornblende-almandine |
| 47-81 | 1.80 | 4.15 | 11.57 | altered pigeonite(?)quartz diabase |
| 48-309 | 1.80 | 4.10 | 11.59 | altered olivine diabase |
| 48-480 | 1.80 | 4.12 | 11.56 | altered olivine diabase |
| 48-533 | 1.80 | | | altered pigeonite-quartz diabase |
| 48-446 | 1.80 | 4.15 | 11.64 | granitized gneiss |

H. B. Wiik of Helsingfors, Finland, analyzed garnet from specimen 48-305 and reported:

| | |
|---|---|
| $SiO_2$ | 37.17 |
| $TiO_2$ | 0.20 |
| $Al_2O_3$ | 16.84 |
| $Fe_2O_3$ | 5.70 |
| FeO | 28.51 |
| MnO | 3.41 |
| MgO | 5.53 |
| CaO | 1.75 |
| $Na_2O$ | 0.32 |
| $H_2O^+$ | 0.20 |
| $H_2O^-$ | 0.00 |
| $Cr_2O_3$ | 0.00 |
| S | 0.07 |
| | 99.70 |
| $-S = 20$ | 0.03 |
| Total | 99.67 |

This chemical analysis indicates that this garnet is almandine, although its index, cell edge, and specific gravity correspond with published data for spessartite.

### *Causes of Metamorphism*

The gneisses in the Grenville subprovince had reached equilibrium and culmination of metamorphism before emplacement of granite 2. Its emplacement could not have been the cause of metamorphism because equilibrium assemblages in the amphibolite facies suggest that (a) white mica and potash feldspar of granite 2 are not members of any equilibrium assemblage in the gneisses, (b) the plagioclase in the gneisses adjacent to areas of granitization was not in equilibrium with the medium that formed granite 2 and reacted with this medium, and (c) hornblende is widespread in the gneisses but does not occur in granite 2.

Considerable areas of granite 1 in the biotite zone are only slightly deformed. Rocks in the amphibolite facies were more deeply buried (as pointed out in the section on Direction of Tectonic Transport) and were more highly deformed than the rocks to the west. They were subjected to higher temperatures and pressures and to greater deformation than rocks in the biotite zone. Considering the factors, temperature and deformation (shearing), one must ascribe importance to shearing in the development of the amphibolite facies. The olivine diabase in the Grenville sub-province, under the influence of shearing, tended to attain the same mineral assemblages found in the much older gneisses, while parts of them remain

unaltered. Shearing was the controlling factor in the small amount of recrystallization in the Gowganda formation. Unsheared material shows no sign of recrystallization, and degree of recrystallization increases with intensity of shearing. The chlorite in the Gowganda formation recrystallized to larger grains, and a leucoxene-like substance takes the place of the finely divided dusty material. Biotite with an appreciable $2V$ has formed at the expense of chlorite. At Rankin Lake very fine needles, apparently amphibole, have developed. However, it should be emphasized that at the main zone of faulting and elsewhere rocks have a cataclastic, texture and hence although deformed were not later recrystallized. Along some of these faults retrograde metamorphism has taken place.

Rocks in the biotite zone in the Temiskaming sub-province are largely undeformed; hence a transfer of heat to this area was involved if the heat for this metamorphism was produced mechanically elsewhere. All three factors increase southeastward: depth of burial, degree of deformation, and grade of metamorphism. Hence, it seems best to assume that both depth of burial (with increase in temperature and pressure) and deformation were important factors, although deformation was probably the more important, whether by production of mechanically generated heat or catalytic action, or both.

An extremely speculative, old explanation for structural and metamorphic features similar to those described perhaps merits mention here. Great areas of the Grenville sub-province have flat foliation and lineation (Pl. 4, no. 30) indicating horizontal movement. This could be due to movement of the earth's crust over its interior. The deep-seated area which underwent horizontal movement was later carried upward along steeply dipping faults such as occur at the contact of the sub-provinces. This explains the accommodation for the great increase in dimensions, in some cases over 300 per cent, parallel to $L_1$ as seen in elongation of mineral grains and boudinage. It is difficult to visualize how this lengthening was accommodated if it was not distributed tangentially to the earth's crust.

## Granitization

### *General*

Granite 1 of the Temiskaming sub-province has zoned feldspars, inclusions, and a massive appearance characteristic, according to some, of a granite emplaced at intermediate depth. Granite 2 in the Grenville sub-province is, on the other hand, a deep-seated granite showing the following features which suggest granitization:

(1) Mineral composition approaching that of the country rock.
(2) Indistinct and disconnected layers of granite parallel to the foliation of the country rock and seldom cutting across it. All gradations between granite and gneiss. Poorly defined contacts between large areas of granite and gneiss.
(3) Microscopic veinlets of quartz and potash feldspar along pre-existing fractures in the country rock.
(4) Quartz and mica fabrics of "undeformed" granite similar to those in the enclosing gneiss.
(5) Clusters of apatite or sphene grains along *s*-surfaces marked by parallel biotites.
(6) Replacement of plagioclase by potash feldspar.
(7) Retention of relict textures and structures as recrystallized alteration products in the plagioclase of granite 2.

The time interval after emplacement of granite 1 (during which it was altered and later metamorphosed), before vast quantities of $K_2O$ and $SiO_2$ were introduced to form granite 2 eliminates the possibility that the two granites originated during the same period of igneous activity.

### *Granitization of Granite 1 in Temiskaming Sub-province*

The most important process was the introduction of $K_2O$ which reacted with the plagioclase to form microcline and epidote (usually in veinlets). The potash feldspar replaced the plagioclase of granite 1 as can be seen from irregular patches and zones of microcline parallel to the cleavage of the replaced plagioclase.

and small islands of unreplaced feldspar with the same optical orientation as larger adjacent grains. During this replacement the alteration products of the plagioclase were in part expelled. Plagioclase grains adjacent to this replacement have recrystallized to clear fresh grains. Chlorite, and epidote from the previous alteration of the plagioclase, and the epidote formed by granitization, remained practically unreplaced. The white micas inside the plagioclase decreased in number and increased in size, while epidote and zoisite decreased in size. All three minerals recrystallized to more regular grains with sharper boundaries against the host plagioclase.

### *Granitization of Gneisses in Grenville Sub-province*

The most essential feature of the granitization in the Grenville sub-province was the replacement of plagioclase by potash feldspar. The minerals produced by granitization were with some exceptions the same as those in the Temiskaming sub-province. These minerals in probable paragenetic sequence from earliest to latest are: sphene, apatite, magnetite (microcline, orthoclase, epidote formed simultaneously by the same reaction), white mica, quartz, and carbonate. The sequence of minerals encountered from a slightly granitized area to one in which granitization was complete probably is: apatite, sphene (in rounded grains), allanite, white mica (produced from biotite), potash feldspar, and quartz.

Quartz and potash feldspar are the only abundant minerals produced by granitization. The potash feldspar is always accompanied by later quartz. However, quartz occurs in veinlets unaccompanied by any of the other minerals. The quartz has replaced the potash and plagioclase feldspars in preference to other minerals, and micas and epidote remain largely unreplaced. Sphene and apatite occur singly or together and in many places where there has been no introduction of quartz or $K_2O$. Apatite is by far the most widespread of the minerals produced by granitization; it seems to have heralded the approaching granitization. In some instances it has replaced epidote, indicating that the introduced $P_2O_5$ reacted with epidote. In the Temiskaming sub-province apatites adjacent to biotite are in some cases surrounded by very small sphenes, suggesting that titanium in the biotite reacted with the apatite-forming medium to produce sphene.

### *Comparison of Granitization in Sub-provinces*

Granitization caused recrystallization of minerals in the chlorite zone in Granite 1 (Temiskaming sub-province). In the amphibolite facies in the Grenville sub-province granitization had the opposite effect. Minerals present broke down as explained under metamorphism. The plagioclase became progressively more albitic with increase of potash feldspar. This feature is also suggested by Wahl's (1950) plagioclase determinations and Rosiwal analyses for the Cawatose map area. Reaction of the granitizing medium with biotite produced white mica; this did not take place in the Temiskaming sub-province. The white mica is usually intergrown parallel to the biotite although in a few cases it cuts across it. Another feature not seen in the Temiskaming sub-province was carbonate pseudomorphic after epidote in Granite 2. In the Grenville sub-province hornblende has altered to mica under the influence of granitization. This was not noticed in the Temiskaming sub-province.

## Conclusions

East of Lake Temagami the Temiskaming and Grenville sub-provinces are in fault contact along a northeast-trending, eastward-dipping, complex zone of faulting; successive northeast segments of the contact lie to the right. These faults have been active many times and have tended to obscure the relationship of the rocks along this contact. The Grenville sub-province was the active block during this faulting. Later northwest-trending, steep-dipping faults followed tension joints parallel to the earlier northwest movement on northeast striking faults.

The emplacement of the granite, metamorphism, and deformation in the Grenville sub-province all postdate the Archean rocks in the Temiskaming sub-province. Some of the

gneisses in the Grenville sub-province probably resulted from metamorphism of granite of the same age as the granite in the Temiskaming sub-province. The biotite isograd in the Temiskaming sub-province is parallel to the contact between the sub-provinces; the grade of metamorphism increases eastward. The granite in the Grenville sub-province is markedly different from the granite of the Temiskaming sub-province chiefly in that it is fresher and much higher in $K_2O$, and hence belongs to an entirely different and separate petrographic province.

Both granites are pre-Cobalt and probably post-Temiskaming (Todd, 1925, p. 9) and therefore not Laurentian. The granite in the Temiskaming sub-province is usually referred to as Algoman in age. The granite in the Grenville sub-province is therefore post-Algoman and pre-Cobalt and occupies a position in the geological column where no granite has previously been reported.

Only a very narrow zone along the contact of the sub-provinces has been examined. To obtain a complete picture a much wider zone should be studied. In fact, a feature as important as this contact should receive more attention than it has been given in the past.

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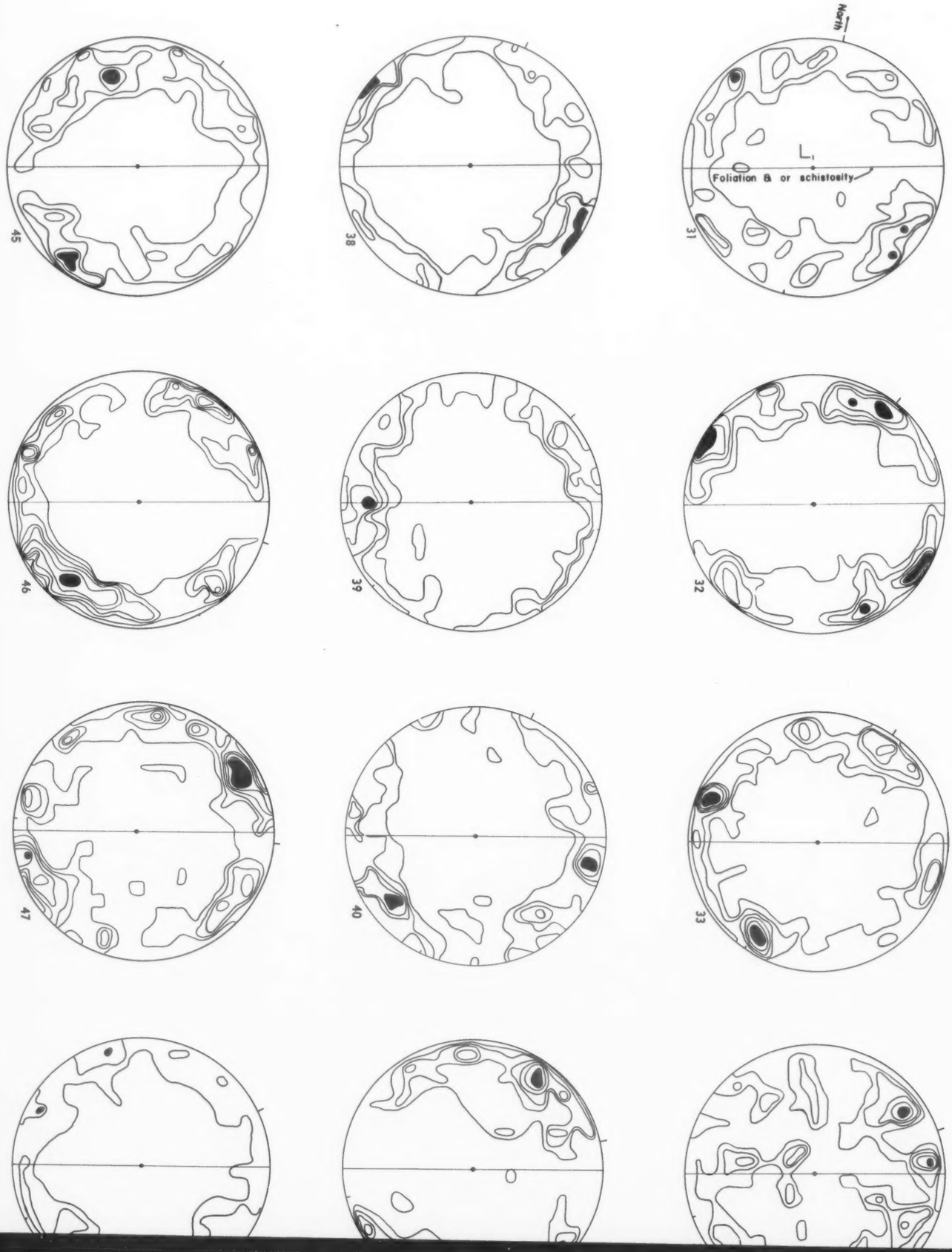
Elk Lake, Northern Ontario, Canada

Manuscript Received by the Secretary of the Society, April 13, 1953

| No. | Specimen No. | Mineral | Contours % | Rock | Area No. on Plate 2 | Location |
|---|---|---|---|---|---|---|
| 31 | 49-233 | 277 quartz | 4, 3, 2, and 1 | Gowganda graywacke | 5 | Cross Lake |
| 32 | 49-222 | 200 quartz | 5, 4, 3, 2, and 1 | Granite 2 | 5 | Cross Lake |
| 33 | 48-208 | 125 quartz | 6, 5, 4, 3, 2, and 1 | Gneiss | 3 | Hangingstone Lake |
| 34 | 49-186 | 100 quartz | 5, 4, 3, 2, and 1 | Amphibolite boudin | 6 | Marten River |
| 35 | 48-330 | 330 quartz | 6, 5, 4, 3, 2, and 1 | Granite 2 | 6 | Red Cedar Lake |
| 36 | 48-413 | 100 quartz | 5, 4, 3, 2, and 1 | Granitized gneiss | 6 | Highway 64 |
| 37 | 48-393 | 160 quartz | 5, 4, 3, 2, and 1 | Gneiss | 6 | Marian Lake |
| 38 | 48-416 | 277 quartz | 5, 4, 3, 2, and 1 | Sheared granite 2 | 4 | East of Panet Lake |
| 39 | 48-444 | 160 quartz | 5, 4, 3, 2, and 1 | Gneiss | 4 | East of Milne Lake |
| 40 | 48-462 | 200 quartz | 5, 4, 3, 2, and 1 | Sheared granite 2 | 4 | Milne Lake |
| 41 | 48-468 | 235 quartz | 6, 5, 4, 3, 2, and 1 | Sheared granite 2 | 4 | North of Milne Lake |
| 42 | 48-43 | 200 quartz | 4, 3, 2, and 1 | Gowganda graywacke | | Jumping Caribou Lake |
| 43 | 49-2 | 220 quartz | 5, 4, 3, 2, and 1 | Lorrain arkose | | Rabbit Lake |
| 44 | 49-82 | 255 quartz | 4, 3, 2, and 1 | Gowganda graywacke | | S.E. Bay Rabbit Lake |
| 45 | 49-91 | 273 quartz | 4, 3, 2, and 1 | Gowganda graywacke | 1 | Ross Lake |
| 46 | 49-95 | 105 quartz | 6, 5, 4, 3, 2, and 1 | Granite | 1 | Burwash Lake |
| 47 | 49-102 | 160 quartz | 6, 5, 4, 3, 2, and 1 | Gneiss | 1 | Maxam Lake |
| 48 | 49-65 | 445 quartz | 3, 2, and 1 | Lorrain arkose | | East of Rabbit Lake |
| 49 | 49-65 | 120 poles to quartz deformation lamellae | 9, 5, 3, 2, and 1 | Lorrain arkose | | East of Rabbit Lake |
| 50 | 49-102 | 86 *Y*'s of epidote | dotted contours 10, 7, 5, 3, and 1 | Gneiss | 1 | Maxam Lake |
| | 49-102 | 86 *X*'s of epidote | dotted contours 5, 4, 3, 2, and 1 | | | |
| 51 | 48-393 | 108 *Y*'s of epidote | dotted contours 15, 5 3, and 2 | Gneiss | 6 | Marian Lake |
| | 48-393 | 108 *X*'s of epidote | dotted contours 5, 4, 3, 2, and 1 | | | |

These diagrams are all perpendicular to lineation $L_1$. Schistosity or foliation is indicated by the solid line, and north by the radial line projecting from the outside of the diagrams.



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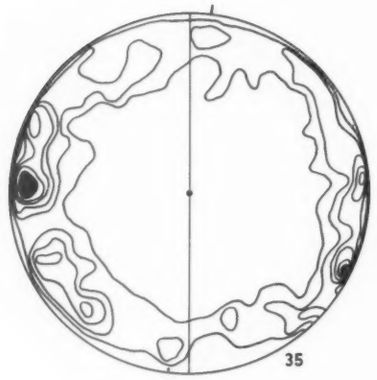
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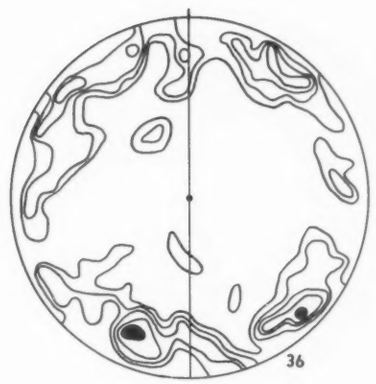
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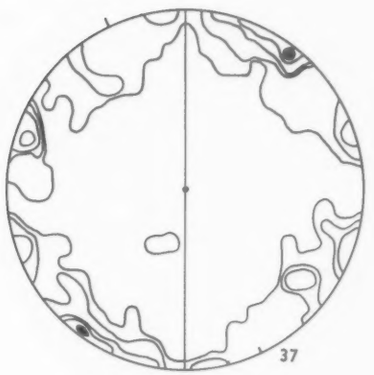
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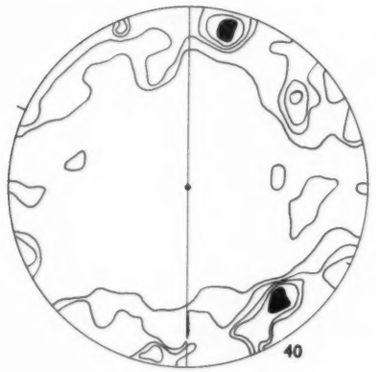
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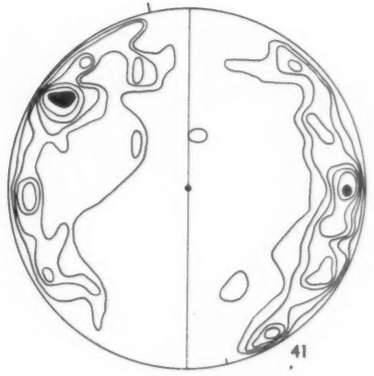
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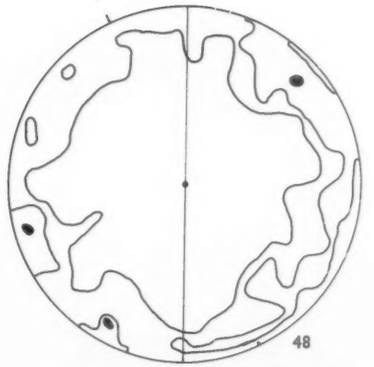
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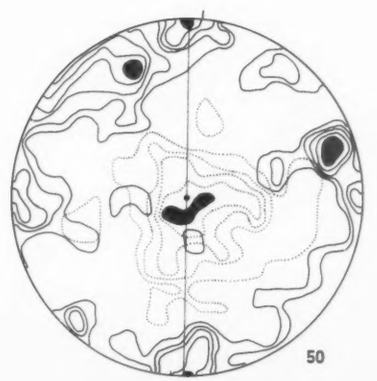
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FABRIC DIAGRAMS OF QUARTZ AND EPIDOTE

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# FORMATION OF EPIDOTE IN TANGANYIKA

By J. R. Harpum

## Abstract

Minerals of the epidote group in Tanganyika are commonly produced during cataclasis of lime-bearing rocks; invariable accession of the hydroxyl radical during dislocation assists their formation. Epidote minerals are not produced from plagioclase by the action of water, even at high temperatures and pressures; shearing stress or the presence of other material is required. Saussuritization follows, and is often a result of, uralitization. Metasomatism, progressive metamorphism, and granitization of lime-bearing rocks can give rise to epidote minerals. Epidote and clinozoisite are stable in the greenschist facies and the albite-epidote amphibolite facies, but are unstable in the almandine-diopside amphibolite subfacies of the amphibolite facies. Zoisite is unstable in the greenschist facies, but is stable in the albite-epidote amphibolite facies and throughout the amphibolite facies. Lime metasomatism, which is responsible for epidotization, is largely a result of displacement by the alkalies of lime from lime-bearing minerals. Epidotization often accompanies ferruginization and sulfide mineralization. Much epidote in the anatectic granites is the result of contamination; primary magmatic epidote is unlikely to occur. The epidote minerals form between 300°C and 500°C. High hydrostatic pressures favor the formation of zoisite. Under conditions of shearing stress monoclinic epidotes are formed. The epidotes are not stress minerals; under metasomatic conditions they are frequently formed without stress.

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## Introduction

In the pre-Karroo crystalline rocks of Tanganyika, members of the epidote group[1] of minerals are common. Epidotization is so extreme in some metamorphic and granitic terrains that numerous occurrences of almost pure epidotic rocks are known. Such rocks have recently been redefined by Flawn (1951, p. 775) as epidotites[2], a term intended to be without genetic significance. The genesis of the epidotites is often difficult to assess, especially if the replacement of the original rock by epidote is far advanced. It is in areas where this replacement is incomplete or arrested that clues to the origin of these rocks may be found.

The principal environments of epidote in Tanganyika are (1) dislocated lime-bearing

[1] Epidote here refers to the ferriferous monoclinic member of the group (=pistacite), epidote series to the isomorphous series clinozoisite-pistacite, and epidote group to the members of the epidote series and zoisite. The term epidotization refers to the production of members of the epidote group by lime metasomatism.

[2] Flawn also uses this term for rocks composed entirely of zoisite or clinozoisite.

rocks, (2) metasomatized or metamorphosed lime-bearing rocks, (3) rocks that have suffered lime metasomatism, and (4) acid magmatic rocks contaminated by calc-silicate xenoliths.

Flawn (1951, p. 774–775) has enumerated two other environments—calc-silicate rocks which have undergone contact-thermal metamorphism, and the products of direct crystallization from water-rich magmas. In Tanganyika no straightforward thermal metamorphism of calc-silicate rocks has been reported; where intrusions have influenced such rocks there is almost invariably some indication of transfer of material in addition to the purely thermal effects. Similarly, pyrogenic epidote has not been recognized in the Territory, although contamination has formed epidote in acid magmatic rocks.

The dominant factor in the production of epidote is, of course, the activity of the calcium ion. The most common calcic component in the crystalline complexes of the Territory is the anorthite molecule of the plagioclase feldspars, although minerals such as hornblende, diopside, and augite are fairly common. The lime released by the breakdown of these lime silicates may be fixed in such phases as members of the epidote group, tremolite-actinolite, calcite, scapolite, fluorite, apatite, and sphene. The widespread occurrence of the epidotes indicates the frequency with which the physico-chemical conditions favorable to epidote formation are satisfied, which in turn suggests that epidote may be formed over a wide range of conditions.

### Acknowledgments

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### Dislocation of Lime-Bearing Rocks

Cataclasis without some accession of the hydroxyl radical is rare. Without chemical analyses this contention is difficult to justify, but characteristic cases indicate this trend. Dislocation of rocks of the granulite facies of metamorphism, rocks which are typically low in their hydroxyl content (Groves, 1935, p. 193), and also of completely consolidated basic and ultrabasic igenous rocks which have not been affected by deuteric activity, invariably results in an accession of the $(OH)^-$ ion from a source apparently outside the rocks. Such primary minerals as labradorite, diopside, augite, hypersthene, garnet, and olivine, on dislocation, change to hydroxylate phases, clinozoisite, epidote, tremolite-actinolite, hornblende, chlorite, and serpentine.

There are two possible sources of this acquired hydroxyl component: either (1) water was already in the rock as intergranular films, in metastable equilibrium with the hydroxyl-free phases, or (2) an influx of water has invariably occurred during episodes of dislocation. This availability of the hydroxyl radical during dislocation is clearly of prime importance for the formation of the hydroxylate epidote minerals.

The writer has noted the dislocation of granulites and charnockitic rocks with the formation of hydroxylate minerals in Ukinga, the Bundali Hills, the Pare Mountains, and the Morogoro District[3]. At an early stage in the dislocation of anorthositic rocks in Ukinga, granulation alone took place, but once the cataclasis proceeded beyond a certain point, the hydroxyl radical manifested itself in clinozoisite, the formation of which would have been precluded but for this influx. Mylonization of a charnockite from the Mpwapwa area (Temperley, 1938, p. 46) has formed albite and much fine-grained epidotic material from the original pyroxene and plagioclase. Dislocation, with the production of hydroxylate phases, is also common in the unmetamorphosed dolerite dikes of Tanganyika, as well as in some of the larger gabbroic and noritic bodies of Ukinga and Upangwa.

The production of epidote in all these cases in rocks otherwise devoid of hydroxylate phases, implies the addition of the $(OH)^-$ ion, supporting the view that dislocation without hydroxylation is rare.

The release of lime by shearing has been demonstrated by Cooke (1927). In Tanganyika, dislocation of rocks containing plagioclase or

[3] *See* Fig. 1 for localities of places mentioned in the text.

lime silicates frequently results in the release of lime which becomes fixed as epidote.

Some of the crystalline limestones or marbles, such as occur in association with the granulite-grade product. Although dislocation of eclogites without metasomatism has not been observed, cases of cataclasis of pyroxene granulites and other rocks of the granulite facies with con-

FIGURE 1.—MAP OF TANGANYIKA SHOWING MAIN LOCALITIES

charnockitic complex of the Territory, contain aluminous impurities. From them during metamorphism grossularite, spinel, phlogopite (with potash), corundum (rare), and anorthite[4] may form. Dislocation of limestones carrying these impurities could result in epidote but this has not been recorded in the Territory.

Dislocation of high-grade, lime-bearing, metamorphic rocks such as eclogites, pyroxene granulites, basic charnockitic rocks, and anorthositic or calcic granulites, nearly always results in the formation of epidote as a retrocomitant production of epidote are common. On dislocation diopside changes first to actinolite and then to chlorite. In the second stage there is an accession of alumina and a release of lime. Clarke (1924, p. 411) noted that the lime released during chloritization of lime-bearing ferromagnesian minerals becomes fixed in epidote; alumina for both epidote and chlorite is derived from the breakdown by stress of the anorthite molecule:

$$4CaAl_2Si_2O_8 + H_2O\ (+\ shearing\ stress) \longrightarrow 2Ca_2Al_3(SiO_4)_3(OH) + 2SiO_2 + Al_2O_3$$
(*clinozoisite*)

Although this reaction does not satisfy the volume law (Turner 1948, p. 121), in this

[4] Anorthite has not been found in the crystalline limestones of Tanganyika but Stockley (1948, p. 13) records labradorite from a crystalline limestone of the Igawa area in the Njombe District.

case other changes proceeding at the same time eliminate the deficiency in volume. The appearance of epidote in the assemblage under such conditions is critical for the facies designation. The writer's experience in Tanganyika has shown that diaphthoresis or retrograde metamorphism by dislocation alone does not produce epidote in a calc-silicate rock until the albite-epidote amphibolite facies is reached; epidote formation by dislocation coincides with the appearance of albite and blue-green, alumina-poor hornblende.

In Ukinga, dislocation of anorthositic granulites free from iron-bearing minerals results in the formation of clinozoisite. On the other hand, dislocation of granulites containing schlieren and irregular masses of hypersthene causes slight transfer between the dark and light parts of the rock; in marginal zones the hypersthene changes to tremolite and the labradorite to epidote.

The behaviour of epidote in metamorphic facies is most clearly demonstrated by the effects of cataclasis on the amphibolites of Ukinga. The writer has distinguished two cataclastic series, both resulting in marked diaphthoretic changes:

*1: Gneiss → mortar gneiss → granulated gneiss → schist. Typified by epidote and titaniferous magnetite.*

*2: Gneiss → flaser gneiss → phyllonite. Typified by clinozoisite and sphene.*

The metamorphic facies changes in both series involve retrogression from the amphibolite facies, first to the albite-epidote amphibolite facies and finally to the green-schist facies. In the first series, the assemblage hornblende, oligoclase, quartz, and titaniferous magnetite breaks down, first to blue-green (alumina-poor) hornblende, albite, quartz, epidote, and titaniferous magnetite, and then to actinolite, chlorite, epidote, albite, quartz, and titaniferous magnetite. In the second series, where the onset of the dislocation is more rapid and more intense, there is retrogression, first to the assemblage actinolite, albite, quartz, clinozoisite, sphene, and magnetite, and then to chlorite, albite, quartz, clinozoisite, ilmenite, magnetite, and possibly epidote. In this last stage the sphene breaks down to ilmenite and a very fine-grained, unidentified, epidotic mineral.

In the second cataclastic series there is no segregation of epidotic material into discrete masses; in the first series the albite-epidote amphibolites frequently show small lenticular segregations of epidote. These epidote segregations occur in an albite-epidote amphibolite which has developed a banding as a result of metamorphic differentiation induced by dislocation. The rock contains hornblende-rich bands with a little albite, quartz and iron ore, and bands rich in quartz, albite, and small epidote lenses. This segregation of epidote within the felsic bands supports the view (Schmidt, 1932, p. 183–187) that, during dislocation, minerals such as epidote, which are not readily deformed by gliding, tend to be mechanically segregated as the more plastic phases, quartz and albite in this case, begin to flow. While this process leads to small-scale epidotite formation, the writer doubts that it could take place on a scale large enough to produce large epidotite bodies.

The differences in the two cataclastic series are mainly due to the intensity of dislocation. The most susceptible mineral to the cataclasis is oligoclase, but whereas in the first series the breakdown of the plagioclase takes place at the same time as that of the other minerals, in the second series it occurs prior to that of the iron-bearing minerals, resulting in the production of clinozoisite rather than epidote. The excess of lime, once the available alumina from the hornblende is used up, produces sphene, often as a rim surrounding the titaniferous iron ores. The formation of sphene is precluded in the first series by the absence of any free lime, once the iron and alumina from the hornblende are fixed as epidote.

During the dislocation of acid gneisses in Ukinga, even where cataclasis is clearly related to Series 2, the tendency to form clinozoisite and sphene is much less than in the more calcic rocks. Similarly the formation of clinozoisite in granitic rocks by dislocation alone is not known. This absence is probably related to the relatively low $CaO:Fe_2O_3$ ratio in the acid gneisses, compared to that for the amphibolites. Therefore, the typical change observed in a quartz-oligoclase-microcline-biotite-magnetite assemblage, on dislocation, is to quartz-albite-microcline-muscovite (or sericite)-chlorite-magnetite-epidote. The change from biotite to chlorite releases iron for the formation of epidote; the generation of sericite from microcline drains the available alumina, precluding the formation of clinozoisite. The production of epidote during the change from oligoclase to albite is common. This type of epidote formation is reported in rocks from the Eastern Lupa Goldfield (Teale and Oates, 1935, p. 17). The writer noted lenticular segregations of epidote in these rocks; these segregations were formed by metamorphic differentiation as a result of dislocation (Schmidt, 1932), in much the same way as those observed in albite-epidote amphibolites of Ukinga.

Dislocation of gabbros, dolerites, and basalts follow trends similar to those in basic gneisses. Products of cataclasis of both these groups are so

similar it is difficult to decide the genesis of a schist where both groups occur. The schistose amphibolites, associated with banded ferruginous quartzites a few miles southwest of Dodoma, belong to the albite-epidote amphibolite facies. The mineral assemblage is blue-green hornblende, epidote, quartz, albite, and sphene. These rocks are usually considered sheared basic volcanics. Epidote is also consistently formed in shear belts in gabbros and dolerites throughout the Territory.

## Action of Water or Water Vapor on Lime-Bearing Rocks

The reactions involving water or water vapor during episodes of dislocation probably take place at low temperatures. Dislocation without an accession of hydroxyl ion is rare and would result in very little change in rocks, apart from granulation; water, however, acts as a catalyst which promotes or accelerates the appearance of hydroxylate diaphthoretic products, notably epidote. The presence of pure water as an intergranular film in rock material does not, at normal temperatures and pressures, lead to the formation of epidote. Some other factor, such as dislocation, an increase in temperature, or an addition of other material, is required.

Normal weathering of plagioclase feldspars does not lead to epidote formation (Goldich, 1938; Winchell 1951, p. 277, 278, 330) which indicates that some increase in temperature might be required for its production from the anorthite component. Ramberg (1944) has concluded that zoisite (epidote) represents the low-temperature and hydrous equivalent of anorthite. Ramberg (1949, p. 30) has further considered the action of water upon anorthite in potash-free and potash-rich systems. These reactions he represents respectively by:

$$\underset{\textit{anorthite}}{4CaAl_2Si_2O_8} + H_2O \rightleftharpoons \underset{\textit{kyanite}}{Al_2SiO_5} + \underset{\textit{epidote}}{2Ca_2Al_3Si_3O_{12}(OH)} + \underset{\textit{quartz}}{SiO_2}$$

and

$$\underset{\textit{anorthite}}{4CaAl_2Si_2O_8} + \underset{\textit{orthoclase}}{KAlSi_3O_8} + 2H_2O \rightleftharpoons \underset{\textit{muscovite}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\textit{epidote}}{2Ca_2Al_3Si_3O_{12}(OH)} + \underset{\textit{quartz}}{2SiO_2}$$

In each reaction the change from left to right is accompanied by a lowering of the temperature. In Tanganyika no evidence for the formation of epidote in either way has been found. Goldsmith and Ehlers (1952, p. 396, 397) failed to synthesize either zoisite or clinozoisite by subjecting the system anorthite-water ($CaAl_2Si_2O_8 + H_2O$) to varying conditions of temperature and pressure over different periods of time. In experiments at pressures ranging from 1000 bars down to that of the vapor pressure of water at the experimental temperature, they observed that anorthite was stable in the presence of water at temperatures down to 350°C and possibly lower. Below 300°C a hexagonal polymorph of anorthite, not known in nature, was consistently produced, but no zoisite. Rosenqvist (1952, p. 50–51) produced a mineral believed to be zoisite by the reaction of calcium hydroxide with plagioclase and water; this discovery is not yet confirmed.

Winchell (1951, p. 278) regards saussuritization a product of dislocation, but the evidence from Tanganyika shows that this type of anamorphism is as characteristic of metasomatic changes as it is of cataclastic changes. Saussuritization, producing albite or oligoclase, zoisite, epidote, calcite, sericite, actinolite, and other calc-silicates, is frequently a product of autometasomatism or deuteric activity in an igneous rock. Deuteric changes result from the addition of water alone (Stewart, 1947, p. 483; Oliver, 1951, p. 428). The epidote mineral may occur in a variety of textural growths, even in the same rock. Fine-grained, granular aggregates and large, irregularly shaped grains are equally common. Epidote has also been observed as a mass of small needles, laths and prismatic crystals set in plagioclase.

Examinations of some of the olivine-free basic igneous masses of Ukinga, the Lupa, and Nyabuyonza show that the first indication of deuteric activity is uralitization. This is also true of the altered dolerites scattered throughout the central part of the Territory.

During uralitization, there is accession of water (Oliver, 1951, p. 428) while lime, silica, alumina, and iron are released, which react with the plagioclase to form epidote or

clinozoisite. Rosenqvist (1952, p. 38–49) has shown that ferric iron has an important influence in the saussuritization process; the epidote minerals of saussuritized rocks contain a considerable amount of ferric iron. Saussuritization, therefore, follows and is a direct result of uralitization. Clarke (1924, p. 609) remarks that epidotization is the result of a reaction between the feldspars and ferromagnesian minerals of a rock which in extreme cases gives rise to a quartz-epidote rock.

Evidence from Ukinga and Upangwa indicates that deuteric effects in anorthositic rocks, poor in ferromagnesian minerals, are limited. The small amount of material liberated during uralitization is not sufficient to produce more than a very local development of epidote. Similarly, calcic or anorthositic granulites of the charockitic suite observed in Upangwa may show uralitization of their pyroxenes, but in the absence of other chemical components or of dislocation there is only slight epidotization of the feldspars, and that only in immediate contact with the ferromagnesian material.

The action of water alone, therefore, on plagioclase feldspars does not result in the production of epidote, even under high pressures and at elevated temperatures. Turner (1948, p. 121) pointed out that the addition of water to anorthite to produce clinozoisite, quartz, and alumina, results in an equation which does not satisfy the requirements of the volume law. In the case of saussuritization by deuteric action, the change in volume would be slight, and the volume law should apply. The action of water alone upon calc-silicate rocks containing pyroxenes does, on the other hand, result in epidote formation by reaction of the released material with the anorthite component. In Tanganyika, however, epidotites formed in this manner have not been reported.

Another mode of epidote formation not commonly reported in the Territory is that in amygdules in basaltic and andesitic lavas.

A vesicular basalt from the Singida area (Eades and Reeves, 1938, p. 23) shows amygdules infilled with calcite, epidote, and chlorite; the same minerals with the addition of quartz are reported from northern Uwanji in the Njombe district (Stockley, 1948, p. 21) as infillings in an amygdaloidal andesite. Grantham (1928, p. 14) reports epidote with quartz in amygdules in andesitic lavas 45 miles southwest of Iringa.

The formation of epidote in such an environment is almost certainly similar, both chemically and mineralogically, to the deuteric activity observed in basic intrusives. Nearly all the material for the epidote formation is derived from the lava itself; the only material added is water. This water was presumably imprisoned in the amygdules of the lava and subsequently reacted with the surrounding material forming epidote; the alteration is therefore a special deuteric effect. In Glencoe, Scotland, epidote occurs in amygdules with withamite (not yet reported in Tanganyika), the material for the formation of both minerals being derived from the lava itself, in this case an andesite. Even the manganese for the production of the withamite is derived (Hutton, 1939, p. 123) from the breakdown of ferromagnesian minerals in the lava; the introduced material need only have been water.

## Metasomatism and Progressive Metamorphism of Lime-Bearing Rocks

Metasomatism of lime-bearing rocks may result in progressive metamorphism, straightforward replacement without change of metamorphic grade, or diaphthoresis. All three processes can produce epidote often on a scale sufficient to form epidotites. Isochemical metamorphism of rocks of suitable composition also results in epidote formation.

A discussion of the metamorphic behavior of the minerals of the epidote group must include some account of their stability in the various metamorphic facies (Fig. 2). In Tanganyika there is no evidence that either epidote or zoisite is stable in the granulite facies of metamorphism. Where minerals of the epidote group occur in such rocks, they are invariably incipient retrograde or alteration features. Eskola (1952, p. 156) regards the epidote in the Lapland granulites an alteration product. Turner (1948, p. 100) suggests, however, that zoisite can be a mineral of the granulite facies, but in Tanganyika this has not been the case. Minerals of the epidote group have not been observed, except as products of diaphthoresis, in the eclogite facies, although zoisite was

originally described in the eclogite from Saualpe in Carinthia (Dana, 1914, p. 514), and Turner (1948, p. 105) also reports zoisite as a primary mineral in some eclogites. Zoisite may be stable in the almandine-diopside amphibolite subfacies[5], for zoisite and diopside occur in equilibrium in rocks belonging to this subfacies. Where epidote appears in a rock of this subfacies, it seems to be a retrograde feature, although Turner (1939, p. 593, 594; 1948, p. 88) regards it as characteristic of the subfacies. Examples of epidote and diopside in the same rock are numerous, particularly in the Mpwapwa area of Tanganyika (Temperley, 1938, p. 25, 35–41), but further examination of these rocks shows that where both minerals occur together, the rock shows signs of disequilibrium in its assemblage. Gavelin (1952, p. 40) considered that epidote was not in equilibrium with diopside and anorthite in the skarn rocks of the Adak area. Clinozoisite is not known in either the granulite facies or the almandine-diopside amphibolite subfacies in Tanganyika. In the normal rocks of the amphibolite facies (*i.e.*, staurolite-kyanite subfacies), epidote, zoisite, and clinozoisite have been recognized, while in the albite-epidote amphibolite facies all are characteristically developed.

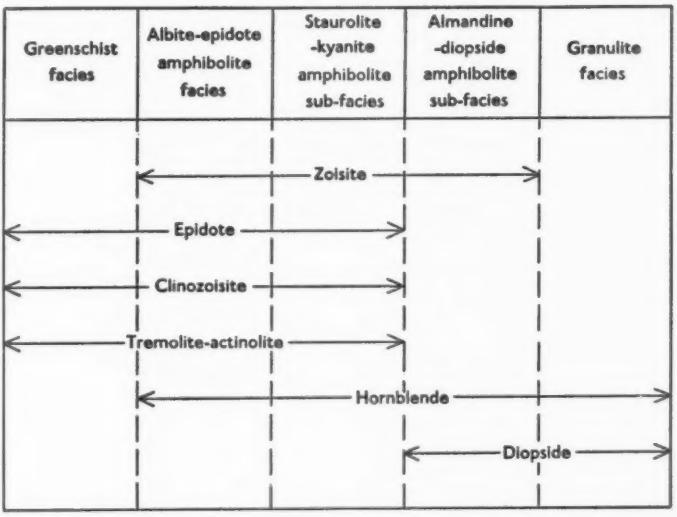


FIGURE 2.—STABILITY RANGES OF CERTAIN CALC-SILICATES IN METAMORPHIC FACIES OF TANGANYIKA

Ramberg (1949, p. 25) has defined the highest part of the albite-epidote amphibolite facies as that in which epidote is in equilibrium with $An_{20-30}$, but the writer finds that in Tanganyika this equilibrium is also possible throughout the staurolite-kyanite subfacies of the amphibolite facies. The upper limit of the albite-epidote amphibolite facies in Tanganyika, marked by the appearance of blue-green hornblende, coincides roughly with the equilibrium of epidote with $An_{10-15}$ (*cf.* Turner, 1948, p. 89.). Both clinozoisite and epidote are well developed in the greenschist facies, but in Tanganyika no zoisite has been noted. Harker (1932, p. 263) describes zoisite-mica schists from Glencoe in Scotland; these, however, lie in the Garnet Grade of the Highland Metamorphic Zones, which is usually regarded as the equivalent of the albite-epidote amphibolite facies and not the greenschist facies. The Glencoe rocks, therefore, probably belong to a facies in which, in Tanganyika, zoisite is stable.

[5] In Tanganyika diopside is absent from the subdivisions of the amphibolite facies except the almandine-diopside amphibolite subfacies. The only subfacies of the amphibolite facies so far recognized in Tanganyika are the staurolite-kyanite and the almandine-diopside subfacies.

The crystalline limestones of this Territory correspond in their original form to the Aldan facies of Korzhinsky (1937, p. 390–391), which is isofacial with the granulite facies. In the marbles of this facies in the Aldan Massif in Siberia, Korzhinsky reports the presence of zoisite, but in Tanganyika no member of the epidote group has been found in such rocks. Under the dry or dipsenic (Rosenqvist, 1952, p. 91) metamorphic conditions of the granulite-charnockitic complex of Tanganyika, limestones resist permeation, whereas under normal conditions of regional metamorphism leading to granitization, in which water is in excess, permeation of the limestones is facilitated, and their alteration may take place on a large scale. Under such conditions epidote may form in the limestones.

This has occurred in the limestones from the Igawa area in southwest Tanganyika (Stockley, 1948, p. 13). The writer noted in this area a true epidotite containing relics of serpentine after forsterite, clearly the product of alteration, by regional metamorphism, of an impure (siliceous) marble. From the same area, Stockley (1927, p. 28) reports epidote formed by the alteration of grossularite (during the retrograde change from the almandine-diopside subfacies to the staurolite-kyanite subfacies of the amphibolite facies).

Metamorphism of impure aluminous and siliceous limestones under conditions of the almandine-diopside amphibolite subfacies produces a diopside-zoisite rock. Diaphthoresis results in an actinolite-epidote rock.

The actinolite-epidote-corundum-spinel and the diopside-zoisite-corundum-spinel rocks of the Longido area in the northern part of the Territory are possibly limestones modified in this manner. Epidote formation is favored more by the presence of calcium-rich rather than dolomitic limestones, although, as is demonstrated in the Longido area, actinolite and epidote are common associates, so that the presence of magnesia in a limestone does not necessarily preclude, but merely limits, epidote formation. In some zones of their outcrop, the Longido rocks are composed of epidote or zoisite alone; these zones are therefore true epidotites. A rock similar to that of Longido, contaning actinolite, epidote, and sphene from the Mpwapwa District, probably has a similar origin. Late stages in the complete migmatization of such rocks may be seen in some of the epidote and clinozoisite rocks of Msagali, near Mpwapwa (Temperley, 1938, p. 41), where epidote, clinozoisite, quartz and plagioclase assemblages, with or without pyriboles, are developed. The change from the diopside-zoisite to the actinolite-epidote assemblages suggests an influx of iron coinciding with the facies change.

The change diopside to actinolite is, according to Turner (1948, p. 88), distinctive of the reversion from the almandine-diopside amphibolite subfacies to the other subfacies of the amphibolite facies; in Tanganyika the change zoisite to epidote under certain circumstances has a similar implication.

Granitization or migmatization of basic igneous rocks, of basic granulites and charnockitic rocks, and of amphibolites, generates free lime which frequently becomes fixed as epidote. During the early stages of the metamorphism, both the igneous rocks and the granulites are converted to amphibolites. The main cause of granitization is an influx of the alkalies in which soda precedes potash. Under these conditions lime is not released on a large scale from amphibolites until potassic influxes begin. In the early stages of the alteration, the breakdown of hornblende to biotite is commonly accompanied by epidote formation in the rock itself.

Epidotization as a result of this action has been observed in the amphibolitic migmatites in the Matumbulu area a few miles south of Dodoma. As metamorphism proceeds, the hornblende disappears leaving biotitic migmatites usually free from epidote. The disappearance of epidote with hornblende, however, does not always take place. Examples of epidote with biotite are also known from the Morogoro area, and Ukinga. The epidote in such cases is invariably associated with the biotite, in clots, so that this type of texture may be considered indicative of a hornblendic parentage. When these rocks become anatectic, epidote disappears, suggesting that the existence of epidote in an acid magma or migma is normally precluded by physico-chemical conditions.

Progressive metamorphism of sheared amphibolitic rocks in Ukinga by potassic influxes has produced mica-bearing, microclinic, potash-permeation augen gneisses. The lime displaced has migrated still father into the amphibolites and by replacement given rise to zones of lenticular epidotites. This secondary lime permeation, although altering

rocks of the amphibolite facies, is, strictly speaking, a low-temperature lime metasomatism affecting comparatively cold rocks.

The nature of the appinites in parts of Tanganyika suggests that they were once limestone (*cf.* Stockley, 1948, p. 16). The stages in the formation of appinitic rocks leading to hornblendic migmatites are visualized as follows:

*magnesian limestone → diopsidic rock → (diopsidic) appinite → hornblendic migmatite.*

Evidence in Tanganyika suggests that the chief factor in these changes is the activity of trondhjemitic material, with a limited potassic influx in the last stage. An unusual rock of appinitic aspect, found at Rungwa, about 120 miles north-northeast of the Lupa Goldfields in south-central Tanganyika, is composed of hornblende and zoisite. In this case the stages in the formation of the rock were probably:

*aluminous limestone → zoisitic rock → "zoisite appinite" (→ hornblendic migmatite ?).*

This closely parallels the suggested mode of formation of the Longido epidotites, where zoisite rather than epidote was produced, and suggests that the true appinites are probably isofacial with the almandine-diopside amphibolite subfacies.

During the autometasomatism of gabbroic rocks from Nyabuyonza in the Karagwe area of northwest Tanganyika and Kungwe Bay on Lake Tanganyika, silicification of the feldspars and not uralitization is the first change observed. This silicification produces the characteristic micropegmatite caused by antipathetic movement of quartz and plagioclase (usually oligoclase). The introduction of silica releases lime, soda, and alumina, which generate saussurite from the anorthite component of the unsilicified plagioclase.

No prehnite has been recorded from Tanganyika but in other parts of the world, prehnite is a common associate of epidote, zoisite, or clinozoisite, in hydrothermal or metasomatic alterations of lime-bearing rocks (Orlov, 1925; Taylor, 1935, p. 122–125; Fraser and Butler, 1938; Watson, 1942). That chemical conditions are not responsible for the production of prehnite rather than iron-poor epidote is clear from their close similarity in composition. In all the cases quoted above, the prehnite formed later than the epidote during a hydrothermal episode, which suggests that it is a lower temperature mineral. Possibly the temperature factor alone determines which shall form. The relation of the epidote group to zeolites such as heulandite has not been investigated in Tanganyika, but temperature again seems the most important factor in determining the mineral species formed.

## Lime Metasomatism

Regional metamorphism and alkali metasomatism of lime-bearing rocks frequently result in lime metasomatism. The displacement of lime and other material by influxes of the alkalies is referred to by Read (1948, p. 193) as the "pint-pot" argument. This displacement in its broadest chemical principles is reversible, while some of the mineralogical changes are likewise reversible. An influx of any material implies that the concentration of this material is being increased in the system in question. Thus in a system, which may be expressed as

$$A + B \rightleftharpoons C + D$$

a marked increase in the concentration (or in other words an influx) of $A$ or $B$ will tend to produce $C$ and $D$, while an increase in $C$ or $D$ will reverse the reaction. The reaction

*Anorthite + soda + silica = albite + lime + alumina*

is, under certain temperature-pressure conditions, reversible. An influx of soda will make the reaction proceed from left to right, but an influx of lime will reverse the action. This increase in concentration of one of the components of a system largely decides the chemical changes during metasomatism; the actual mineral phases produced, however, also depend upon temperature and pressure conditions.

Migmatization of lime-bearing rocks leads to extensive replacement of calcium by sodium and potassium. Gold mineralization in the Territory is frequently accompanied by sericitization and albitization caused by the activity of potash and soda respectively. Common gangue minerals are calcite and epidote; the development of both is clearly related to the alkali influx.

Turner has shown (1948, p. 122) how the action of soda, silica, and water on the anorthite

component of the plagioclase feldspars leads to the formation of clinozoisite and albite:

$$12\,CaAl_2Si_2O_8 + 2Na_2O + 6SiO_2 + 3H_2O \longrightarrow 6Ca_2Al_3Si_3O_{12}(OH) + 4NaAlSi_3O_8 + Al_2O_3.$$

During the formation of augen gneisses by potash permeation, the action of the lime and soda driven out by the incoming potash affects the pre-existing plagioclases in the rock. Alteration of these plagioclases first manifests itself in the formation of minute rods of clinozoisite, which eventually produce a clouded core in the crystals with a clear rim of sodic plagioclase. This texture is the result of an activity similar to that shown in the above equation, and is also characteristic of the early stages of lime metasomatism of granitic rocks.

The replacement of anorthite by albite need not necessarily lead to the formation of epidotic material within the rock itself. The displaced material may produce lime metasomatism elsewhere,

$$2CaAl_2Si_2O_8 + Na_2O + 2SiO_2 \longrightarrow 2NaAlSi_3O_8 + 2CaO + Al_2O_3$$

(*removed in solution*)

The conversion of soda-lime feldspar to secondary muscovite as a result of potash metasomatism is another source of lime. Sericitization of the plagioclase invariably begins before the sericitization of the potash feldspar. This potassic invasion displaces both soda and lime from the plagioclase. The soda often forms clear albite rims around the sericitized core, a phenomenon observed by Anderson (1937, p. 62) and Cheng (1943, p. 142); the lime may either form epidote in the rock itself or migrate giving rise to lime metasomatism. The release of lime from the anorthite component in this manner may be expressed:

$$3CaAl_2Si_2O_8 + K_2O + 2H_2O \longrightarrow 2KAl_3Si_3O_{10}(OH)_2 + 3CaO$$

(*muscovite*)

This reaction represents a decrease in volume of about 11 per cent, but, as changes involving the formation of epidote and other minerals may be taking place at the same time, the requirements of the volume law need not necessarily be satisfied by this equation. The lime released may either cause reactions in the rock itself or may migrate, giving rise to lime metasomatism elsewhere, beyond the zone of sericitization. As the relative amount of potash increases, microcline rather than sericite is produced; this action displaces further lime and alumina:

$$2CaAl_2Si_2O_8 + K_2O + 2SiO_2 \longrightarrow 2KAlSi_3O_8 + 2CaO + Al_2O_3.$$

The production of potash feldspar porphyroblasts, preceded and sometimes succeeded by sericitization, is common in the Territory. Although both lime and soda released by potash permeation are then available for metasomatism elsewhere, the lime usually migrates farther than the soda; under these circumstances albitization is usually preceded by epidotization.

An influx of potash will also cause the formation of biotite from calc-silicates such as diopside, augite, and hornblende. The change from hornblende to biotite commonly generates lime which may calcify adjacent plagioclase feldspars, form epidote, or migrate from the rock. The change may be expressed:

$$5Ca_2(Mg, Fe)_4Al_2(OH)_2Si_8Al_2O_{22} + 3K_2O + H_2O \longrightarrow 3K_2(Mg, Fe)_6Al(OH)_4Si_6Al_2O_{20} + 15SiO_2 + 10CaO + 4Al_2O_3.$$

(*hornblende*) (*biotite*)

In dominantly amphibolitic terrains this breakdown of hornblende may be responsible for quite as much lime metasomatism as the breakdown of the anorthite molecule.

Once lime metasomatism is induced, epidote is a common product (Gavelin, 1932, p. 27, 31). Should the breakdown of the anorthite molecule be effected by carbonate solutions (Turner, 1948, p. 122) then the $CO_2$ in the system will form calcite. Similarly the activity of chlorine, fluorine, phosphorus, and titanium limits the formation of epidote in favor of such phases as scapolite, fluorite, apatite, and sphene. The presence of these components does not, however, preclude epidote formation; for many cases of their association with metasomatic epidote are known. The production of

epidote, even in the presence of carbon dioxide, chlorine, and fluorine is favored by high temperature, when these volatiles become fugitive leaving the remaining constituents to form epidote.

Feldspars in granitic rocks are often pink. This coloration is the result of staining by metasomatic iron oxide introduced along crystal boundaries and cleavage planes. Ferruginization of this type frequently accompanies epidotization; the result is a striking pink and green rock.

Such rocks are particularly well developed around Dodoma, but have also been observed in the neighbourhood of Tabora, Nzega, Iringa, and Njombe, and in the Lake Victoria and Lupa gold fields. A specimen from Luika Mine in the Lupa gold field, brought to the writer's notice by A. M. Quennell, shows a veinlet of epidote 1 mm wide cutting a medium-grained gray granite. For 10 mm on either side of the veinlet the granite is pink. This staining is the result of iron metasomatism along the fissure in which the epidote is now seen. The epidotization and the ferruginization were probably concomitant, but the ferric iron responsible for the staining permeated the wall rock of the veinlet more extensively than the lime responsible for the formation of the epidote.

The accompaniment of epidotization by influxes of iron is further manifested by the frequent association of epidote and pyrite (Gavelin, 1952, p. 25, 27).

In Tanganyika this association is common in the Lupa goldfield (Teale and Oates, 1935, p. 30, 35,38) and the Lake Victoria gold fields (Teale and Oates, 1946, p. 15).

In the presence of sulfur, the tendency to form pink feldspars is considerably reduced, while the epidotic mineral may be deprived of some of its iron content to satisfy the amount required for pyrite formation. Thus alkali influxes, which are themselves frequently accompanied by late-stage sulfide mineralization, may also be preceded by sulfur-bearing fluids which accompany the displaced lime and iron, giving rise to an epidote-pyrite complex. This does not necessarily indicate two distinct episodes of pyrite formation; it may indicate that the sulfur, which originally invaded the rocks with the alkalies, overtook the displaced zone of lime metasomatism.

The rocks usually affected by epidotization in Tanganyika are granites, granitic gneisses, acid dikes, quartz veins, amphibolites, and basic igneous rocks. Most of the epidotites have formed in granites, aplites, and quartzitic rocks. Lime metasomatism produces more epidotites than any other process mentioned. In granitic rocks which have a certain amount of readily available iron, epidote is formed through intense lime metasomatism to which both potash and plagioclase feldspars succumb; the potash and soda are displaced and dispersed by the lime much as the lime was liberated by alkalies. The epidotization in the granitic rocks is usually confined to zones (often structural) or limited areas within the rock mass. Neither the lime nor the displaced soda and potash show any tendency to form fronts; the permeation is either irregular or along structural weaknesses. Although the primary influx of alkalies that initiated the lime metasomatism may have passed as a wave through the rocks, the irregularities and differences in the rock mass cause varying permeation rates of secondary metasomatic activities, so that any fronts are gradually broken up.

During the lime metasomatism of granitic rocks, epidotization first manifests itself as veining, but later the alteration becomes more general. Ferromagnesian minerals and potash feldspar are replaced by the encroachment of epidote grains or grain complexes. Quartz is the last mineral to be replaced, so that a bimineralic quartz-epidote assemblage is the usual product of epidotization of acid rocks. Alteration of plagioclase in most cases produces a fine-grained clouding by epidote of the cores of the feldspars, with the concomitant production of clear sodic rims. The absence of this rim and core texture from a rock containing sodic plagioclase, quartz, and a considerable amount of epidotic material may usually be considered an indication of anamorphism (saussuritization in this case) rather than of lime metasomatism.

In alumina-poor rocks which carry magnesia, tremolite-actinolite is frequently formed instead of epidote as a result of lime metasomatism. Thus, almost pure tremolite rocks have been developed from ultrabasic ortho-

pyroxene-rich bodies in Ukinga and at Kungwe Bay, as a result of the activity of lime-bearing solutions. Lime metasomatism of hornblende may produce epidote (Gavelin, 1952, p. 27, 31), which appears as a granular alteration product along the cleavages of the host mineral (Wooldridge, 1925, p. 243).

Epidotization is to a certain extent selective. Thus, in Ubena, epidotization along the unconformity of a series of phyllites upon granite has a profound effect upon the granite and none on the phyllites. Yet the post-phyllite age of the epidotization has been clearly demonstrated a few miles west of this unconformity in Ukinga, where epidote has formed in schists of the same age as the phyllites. Here the phyllites resisted the lime permeation until they were metamorphosed to schists. That the presence of feldspars is not the controlling factor is shown by the fact that quartz veins in the granite were epidotized, and the schists in only a few cases carried feldspar.

The marked epidotization of quartz veins in the migmatite and granite terrains of Tanganyika is probably the result of structural rather than chemical selectivity. Quartzitic rocks are particularly resistant to permeating fluids, but the fractured nature of most quartz veins may provide an easier passage for incoming fluids than the country rock in which the vein occurs. Using the pint-pot argument, epidotization of quartzitic rocks would be expected to produce silicification elsewhere. While unequivocal evidence of this cannot be advanced, the subsidiary epidote-free quartz veinlets frequently seen emanating from larger epidotized veins may be a result of this activity. As the structural weakness provided by quartz veins probably assists epidotization, a similar argument applies to the numerous epidotized acid dikes found particularly in the gold fields. The almost invariable epidotization of such rocks is clearly a structural rather than a compositional feature, for they frequently occur in epidote-free granitic country rocks with a composition similar to that of the unaltered dikes. True epidotites formed from such rocks are common.

## Contamination of Acid Magmas

Various authorities, including Johannsen (1932, p. 181), have doubted the primary origin of epidote in magmas.

Evidence from Tanganyika is consistent with Johannsen's view, but the presence of epidote in certain areas of the Younger Granites or Lake Granites south, southeast, and east of Lake Victoria cannot be accounted for as the result of late hydrothermal or later metasomatic activity. These granites which are mobilized rocks of anatectic origin, frequently contain xenoliths and even large enclaves of calc-silicate rocks. The typical mineral content of the smaller calc-silicate xenoliths is hornbende, quartz, oligoclase, apatite, epidote, sphene, and iron ore. In the Nzega area, where the writer has studied the Younger Granite in some detail, the process of assimilation was not mechanical but a chemical transfer of material with the surrounding magmatic rock.[6] This mode of assimilation is shown clearly in numerous cases where individual minerals in the xenoliths are considerably finer grained than the same phases in the contaminated granite.

Nockolds (1933, p. 583) concluded that where the breakdown of xenoliths is due to reciprocal reaction between them and the magma, and not to mechanical disintegration, the process produces in the contaminated magma those phases present in the xenolith.

Confirmation of this is found around Nzega, where, with the exception of diopside, all phases of the xenoliths are also present in the surrounding contaminated granite. Epidote, which is a common phase of the xenoliths, is common in the surrounding rock; the absence of any signs of instability shows that this epidote is not the product of later hydrothermal activity. Although, therefore, epidote in such an environment is not strictly a pyrogenic mineral, it may have crystallized in an unconsolidated magma.

Most of the xenoliths observed were lime-rich; they were either calc-silicate rocks caught up by the intruding magma, or enclaves in the complex from which the Younger Granites were derived by anatexis. The field relations at Nzega suggest the latter, for the abundance of migmatitic xenoliths in the Younger Granite and the paucity of lime-rich rocks in the intruded Nyanzian System indicate that these mobilized granites were largely derived from a complex of rocks entirely different from the banded ironstones and acid volcanics of the Nyanzian.

The xenoliths in the Nzega granites, therefore, probably represent relics of larger granite resister

[6] The earliest stage of the breakdown of these enclaves was almost certainly mechanical in part, but at the stage of the assimilation exposed today chemical transfer was dominant.

enclaves (Read, 1951, p. 7) in the complex from which these granites were derived. The epidote within the xenoliths could have formed at several stages. In some of the fine-grained xenoliths there is no variation between the contact with the granite and the core. These xenoliths strongly resemble those amphibolitic rocks, found southwest of Dodoma, that represent the isochemical dislocation metamorphic products of basic volcanics. The formation of the epidote in such xenoliths, therefore, was not necessarily related to the Younger Granite episode. On the other hand some xenoliths show coarse diopsidic cores with a rim of slightly finer-grained hornblende, epidote, apatite, and sphene, indicating a type of diaphthoresis induced by accession of material such as alumina, phosphorus, titania, and iron from the magma. While the metamorphic relics in the main migmatitic complexes of Tanganyika can be isofacial with the almandine-diopside amphibolite subfacies of metamorphism, diopside seems unstable in the environment of the Younger Granites, so that it retrogresses to amphibole and epidote. Epidote so formed is therefore directly related to the Younger Granite episode and is usually more coarse-grained than that formed in the enclaves prior to the mobilization of the Younger Granites. The diaphthoretic formation of epidote and hornblende from diopside-bearing rocks is probably due to permeation and not to dislocation resulting from movement of the surrounding magma.

In mobilized granite in the Singida area (Eades and Reeve, 1938, p. 26, 27, 29, 34, 41) epidote has also formed by contamination. An interesting occurrence of epidote is recorded from Samamba, 10 miles south of Singida (Oates, 1936, p. 15; Eades and Reeve, 1938, p. 41), where xenolithic material, containing melanite, calcite, epidote, orthite, scapolite, pyrite, apatite, and sphene, was noted in a granite. Eades and Reeve believe this contamination is due to limestone, although a considerable amount of material—notably S, Ti, Ce, and possibly Cl and P—would have to be supplied by the surrounding magma, if the limestone is anything like the normal type of crystalline limestone in Tanganyika. The mineral assemblage is chemically remarkable in that it is rich in calcium, fairly rich in iron, and poor in alumina. The epidote showed a 2V of approximately 70°, indicating that it contained over 15 per cent $Fe_2O_3$ (Winchell, 1951, p. 449) and was, therefore, comparatively poor in alumina. The parent rock of such a xenolith may have been an iron-bearing (sideritic, hematitic, or magnetitic) limestone poor in magnesia and alumina; the silica may have been partly original or all introduced. Sideritic and other ferruginous limestones are known around Buhemba Mine, about 25 miles southwest of Musoma on Lake Victoria, in an area where granites similar to those of Singida exist.

The formation of epidote in xenoliths by the action of granitic magma and in granite by the contamination due to such xenoliths does not, insofar as evidence in Tanganyika shows, lead to the formation of epidotites, even though epidote production under these conditions is so common.

Does the epidote in contaminated granites, form in a magma or in a solid rock? The question of pyrogenic epidote has been discussed by several workers (Flawn 1951, p. 773, 774); the believers in primary magmatic epidote maintain that this mode of epidote formation requires a magma with a high water content to account for the low crystallization temperature required. That this content is strictly limited has been indicated by Goranson (1931; 1936; 1937), who concluded that a silicate melt cannot absorb more than 10 per cent of water. This factor alone restricts epidote formation in magmas. Similarly, the comparatively low redox potential even in acid magmas (Goldschmidt, 1943; Mason, 1949, p. 71; Rankama and Sahama, 1950, p. 230, 645) limits the formation of a mineral such as epidote, in which all the iron occurs in the fully oxidized ferric form. Rosenqvist (1952, p. 49) has shown that in addition to a high hydroxyl ion activity, very high ferric ion activity would be required for the formation of magmatic epidote. Mellis (1932), Eskola (1946), and Simonen (1948) indicate the hydrothermal or metasomatic origin of epidotes in rocks such as the helsinkites (formerly regarded as entirely magmatic in origin); this suggests that the epidote produced in the Younger Granites of Tanganyika by contamination did not form in a magmatic environment. Nockolds (1933) does not mention epidote among the products of contamination. This epidote probably formed by diffusion of material from the xenolith, either during the final stages of consolidation when protoclastic textures were being produced in the granite, or immediately after consolidation of the magma before late hydrothermal activity set in. This diffusion was probably taking place throughout the whole event, but under magmatic conditions hornblende would have been formed. The epidote produced by contamination is restricted to the vicinity of the contaminating

xenoliths. The absence of evidence of replacement during this process may be accounted for by the fact that the material moving out from the xenolith producing the epidote is being exchanged volume for volume with material from the host rock which is moving into the xenolith at the same time. Nothing is introduced from outside sources that might produce replacement textures. The exchange of material between the xenolith and the host rock is probably due to a chemical or thermodynamic instability between the two bodies, occasioned by differences in concentration or temperature or both. The formation of epidote in acid magmatic rocks by contamination is thus a special type of lime metasomatism.

A case of the formation of epidotic material by the contamination of a basic igneous rock has been noted by the writer in an intrusion breccia in the Matumbulu area, about 5 miles south of Dodoma. This rock was previously described as lamprophyric (Wade and Oates, 1938, p. 42), but the writer has found it to be of doleritic parentage. Here the reaction between doleritic material and the acid migmatitic country rock produced a plug of clinozoisite-biotite-plagioclase rock carrying pyrite and numerous xenoliths of the comparatively unaltered migmatite. In this case the formation of clinozoisite has been effected, not only by direct reaction with the acid country rock but also by the influx of excess volatiles, chiefly water, during the event. These volatiles also generated epidote veins which cut the contaminated rock and xenoliths, once the whole rock had been consolidated. The epidotes in this intrusion breccia are, therefore, hydrothermal in origin.

## Epidote Stability Fields

The exact figure for the lower temperature limit of epidote formation is not known, but it is probably about 300°C. (Stringham, 1952, p. 662). According to Ramberg (1949, p. 31) formation of epidote in place of the anorthite component is favored by a lowering of the temperature; the highest temperature at which epidote is stable is between 400°C. and 500°C. Rosenqvist (1952, p. 58) has reached similar conclusions concerning the maximum temperature of stability of epidotes.

The formation of epidote from the anorthite component of plagioclase feldspars is also aided by an increase in the Fe/Al ratio (Ramberg, 1949, p. 31). The epidote lattice unlike the feldspar lattice can accommodate $Fe^{3+}$ in place of $Al^{3+}$. In addition, the ubiquity of ferric iron insures the formation of epidote rather than clinzoisite or zoisite. As Winchell (1951, p. 446) has pointed out, when the iron ($Fe_2O_3$) content of zoisite exceeds 5 per cent, the mineral probably reverts to the monoclinic phase. The $Fe^{3+}$—$Al^{3+}$ diadochy in zoisite is thus limited but is more nearly complete in the monoclinic form. The frequent association of metasomatic ferruginization or pyritization with epidotization, observed in Tanganyika, is considered one of the main reasons for the production of epidote rather than zoisite or clinozoisite.

In the almandine-diopside amphibolite subfacies, high confining pressures prevent the reaction between zoisite and iron to form epidote, but in metamorphic facies other than the granulite facies, this change normally takes place in iron-bearing rocks.

There are assemblages where both epidote (or clinozoisite) and zoisite occur in the same rock, but it is suggested that these are unstable. The change from epidote to zoisite may occur as a result of the progression from the staurolite-kyanite subfacies to the almandine-diopside subfacies, while the reverse change is the result of diaphthoresis which may be brought about by permeating iron-bearing solutions. Under normal conditions of deuteric activity, where low temperatures prevail, zoisite is a common product, particularly in the magnesian-rich iron-poor basic igneous rocks. Yet under the low-temperature conditions of formation of the greenschist facies zoisite has not been found. Temperature, therefore, is not the sole control. At the lower limits of their formation prehnite will probably form in the place of all members of the epidote group at similar temperatures. The association of late prehnite with epidote (Fraser and Butler, 1938; Taylor, 1935), with clinozoisite (Orlov, 1925), and with zoisite (Watson, 1942) in other parts of the world supports this view.

High hydrostatic pressures seem to preclude the formation of the monoclinic forms and favor the development of zoisite. The difference in behaviour of the monoclinic and orthorhom-

bic forms to pressure is reflected in a slight difference in their packing indices (Fairbairn, 1943); that of zoisite (6.2) is slightly higher than that of epidote and clinozoisite (6.1). The observations of Ito (1947, p. 231), which confirm those of Waldbauer and McCann (1935), show that zoisite is a mosaic made up of submicroscopic multiple twins of clinozoisite. The replacement of $Al^{3+}$ by $Fe^{3+}$ in the lattice renders the twinning of the epidote series irregular. According to Ito, the intimately twinned zoisite structure is more symmetrical and compact than the normal structure required in the epidote series. This explanation accounts for the differences in the packing indices and in the different reactions to high pressure and shearing stress shown by epidote and clinozoisite on the one hand and zoisite on the other.

The common occurrence of the epidotes in sheared rocks led Harker (1932, p. 149) to class both epidote and zoisite as stress minerals. Turner (1948, p. 98) remarks that under certain conditions, the formation of assemblages with epidote is favored by higher shearing stress and lower temperature than is the formation of assemblages without epidote. In common with a number of other stress minerals, however, epidote often occurs in low-temperature assemblages formed either by diaphthoresis or by low-grade metamorphism, both induced by metasomatism or dislocation. These minerals are just as characteristic of metasomatic episodes which show no sign of contemporaneous shearing stress as of episodes in which shearing stress is important. Evidence from Tanganyika suggests that zoisite cannot exist under the conditions of shearing stress found in the greenschist facies. In the case of the epidote group therefore, the stress and antistress concept of Harker (1932, p. 147) does not apply. Thus, although both zoisite and epidote (or clinozoisite) can form over a fairly wide range of temperatures and pressures, their differing modes of occurrence are governed by the degree of shearing stress, the amount of available iron and alumina, and the limiting effects of high confining pressures. The evidence from Ukinga suggests that formation of clinozoisite rather than epidote during dislocation metamorphism depends to a certain extent upon the nature of the cataclasis. The differences in the mode of occurrence of zoisite and clinozoisite depend on the degree of the $Fe^{3+}$—$Al^{3+}$ diadochy and on the degree of the shearing stress; a high shearing stress favors clinozoisite formation. Clinozoisite formed in the absence of shearing stress may be a ferrian variety of that mineral. The optic angle of clinozoisites in the Matambulu intrusion breccia, where no shearing stress operated during intrusion, is over 80°, indicating that the mineral was approaching epidote. In this particular case a smaller amount of available iron would have resulted in zoisite and a slightly higher amount in epidote. This supports the view of Orlov (1926) and Winchell (1951, p. 446, footnote 144) that zoisite with a content of 5 per cent $Fe_2O_3$ (= $\beta$-zoisite) is monoclinic. Evidence from Tanganyika suggests that there is a range of chemical composition not covered by zoisite and epidote (pistacite); the small gap is filled by clinozoisite and not zoisite. This belief cannot be further justified without chemical analyses and must remain for future research.

The tectonic and chemical fields of stability of the members of the epidote group are summarized in Figure 3. The writer suggests that nowhere do the fields of formation of zoisite, clinozoisite, or epidote overlap (Fig. 3). Occurrences of two of these minerals in the same rock are known in Tanganyika and elsewhere. In these cases, either variations existed in conditions during the formation of the rock, or the rock was the product of at least two events. Disequilibrium phenomena in the metamorphic rocks of Tanganyika indicate that assemblages containing more than one mineral of the epidote group, or containing chemically different members of the same mineral of this group, are not stable. In other words, in equilibrium assemblages different members of the epidote group are mutually exclusive. In the Longido epidotic assemblages for example, rocks containing both zoisite and epidote show a micropegmatitic type of texture caused by the replacement of zoisite by epidote.

Epiditotes are usually formed by lime metasomatism of granitic, quartzitic, and amphibolitic rocks, and of acid dikes, although regional metamorphism of what were probably

impure aluminous limestones in the Longido area and in the Igawa area has produced, in places, almost pure epidotic rocks. The most potent cause of lime metasomatism is the influx of alkalies, particularly potash. Much of this activity takes place during the late hydrothermal stages of migmatic and magmatic episodes. According to Stringham (1952, p. 662), epidote forms under hydrothermal conditions in a fairly restricted physico-chemical field, but its ubiquity during such episodes in Tanganyika suggests either that its field of formation is larger, both physically and chemically, than Stringham indicates or that these particular conditions are very common.

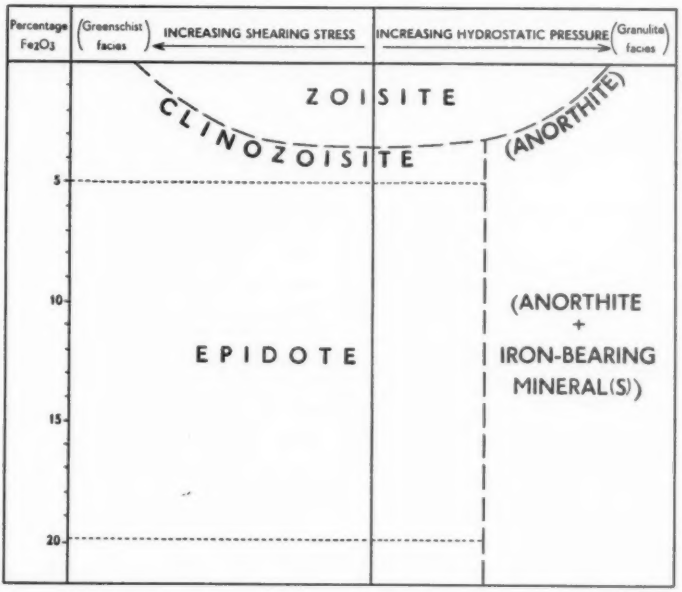


FIGURE 3.—DIAGRAMMATIC REPRESENTATION OF COMPOSITIONAL AND TECTONIC FIELDS OF EPIDOTE GROUP IN TANGANYIKA

During progressive regional metamorphism, in the absence of free iron, the normal member of the epidote group present is zoisite (Kennedy, 1949, p. 50), which is converted to anorthite once the granulite facies is reached, if not before. The change from zoisite to anorthite does not readily take place; the water content of epidote is very strongly bound and is not released at temperatures below 1000°C, under conditions of atmospheric pressure (Smethurst, 1935). The difficulty with which the epidotes part with their combined water may account for the occasional appearance of zoisite in rocks of the granulite and eclogite facies recorded from various parts of the world. In the presence of free iron, during progressive metamorphism, epidote is formed, but it is converted to zoisite in the almandine-diopside amphibolite subfacies, as the first step in the change to anorthite in the granulite facies. If shearing stress upon the system passes a certain limit not as yet ascertained, but probably coinciding with that prevailing in the greenschist facies, clinozoisite and epidote rather than zoisite are formed (Fig. 2).

During regional metamorphism leading to migmatite formation, epidote appears first during the progressive metamorphism of lime-rich rocks, but later disappears at the expense of plagioclase, as sodic influxes begin. Later potassic influxes produce biotite from hornblende, with the formation of epidote, which, however, is seen only if the process is arrested

at this stage. During the early stages of the formation of igneous bodies, epidote does not appear, although prior to consolidation epidote may appear as a result of contamination. Late-stage hydrothermal activity during igneous episodes or toward the close of migmatitic episodes almost invariably results in the formation of some epidote. Most of the epidotes and epidotites in Tanganyika's granitic rocks and acid dikes have formed as a result of late-stage activity of this nature; the late hydrothermal production of the alkalies under these conditions is the chief cause of the lime metasomatism.

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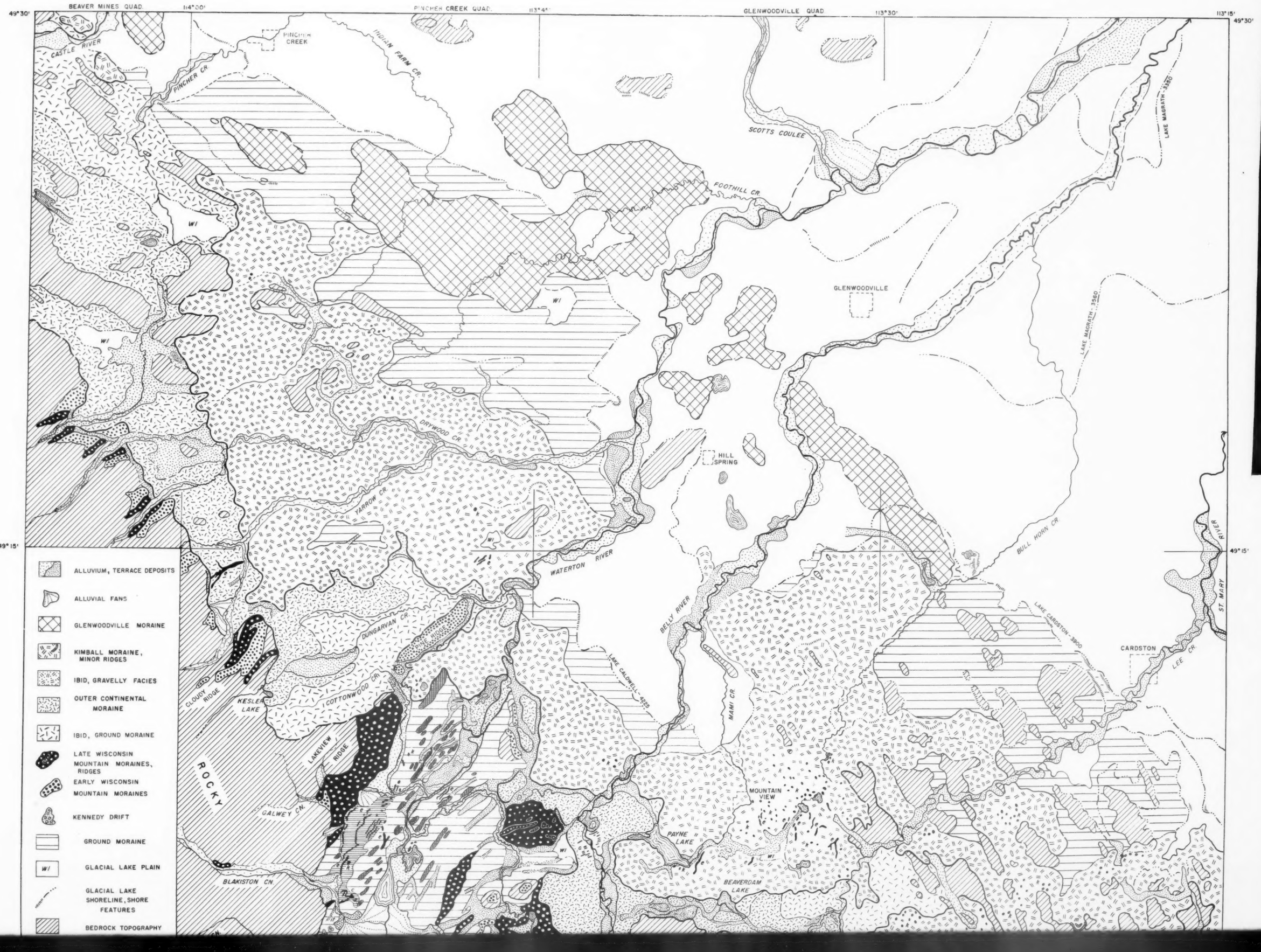
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SCOTTS COULEE
FOOTHILL CR.
LAKE MAGRATH - 3560
GLENWOODVILLE
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DRYWOOD CR.
HILL SPRING
YARROW CR.
WATERTON RIVER
BULL HORN CR.
BELLY RIVER
ST MARY RIVER
LAKE CARDSTON - 3800
CARDSTON
LEE CR.
DUNGARVAN CR.
PINE RIDGE
CLOUDY RIDGE
KESLER LAKE
COTTONWOOD CR.
LAKE CALDWELL - 4225
MANY CR.
LAKEVIEW RIDGE
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GALWEY CN.
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PAYNE LAKE
BEAVERDAM LAKE
ALLUVIUM, TERRACE DEPOSITS
ALLUVIAL FANS
GLENWOODVILLE MORAINE
KIMBALL MORAINE, MINOR RIDGES
IBID, GRAVELLY FACIES
OUTER CONTINENTAL MORAINE
IBID, GROUND MORAINE
LATE WISCONSIN MOUNTAIN MORAINES, RIDGES
EARLY WISCONSIN MOUNTAIN MORAINES
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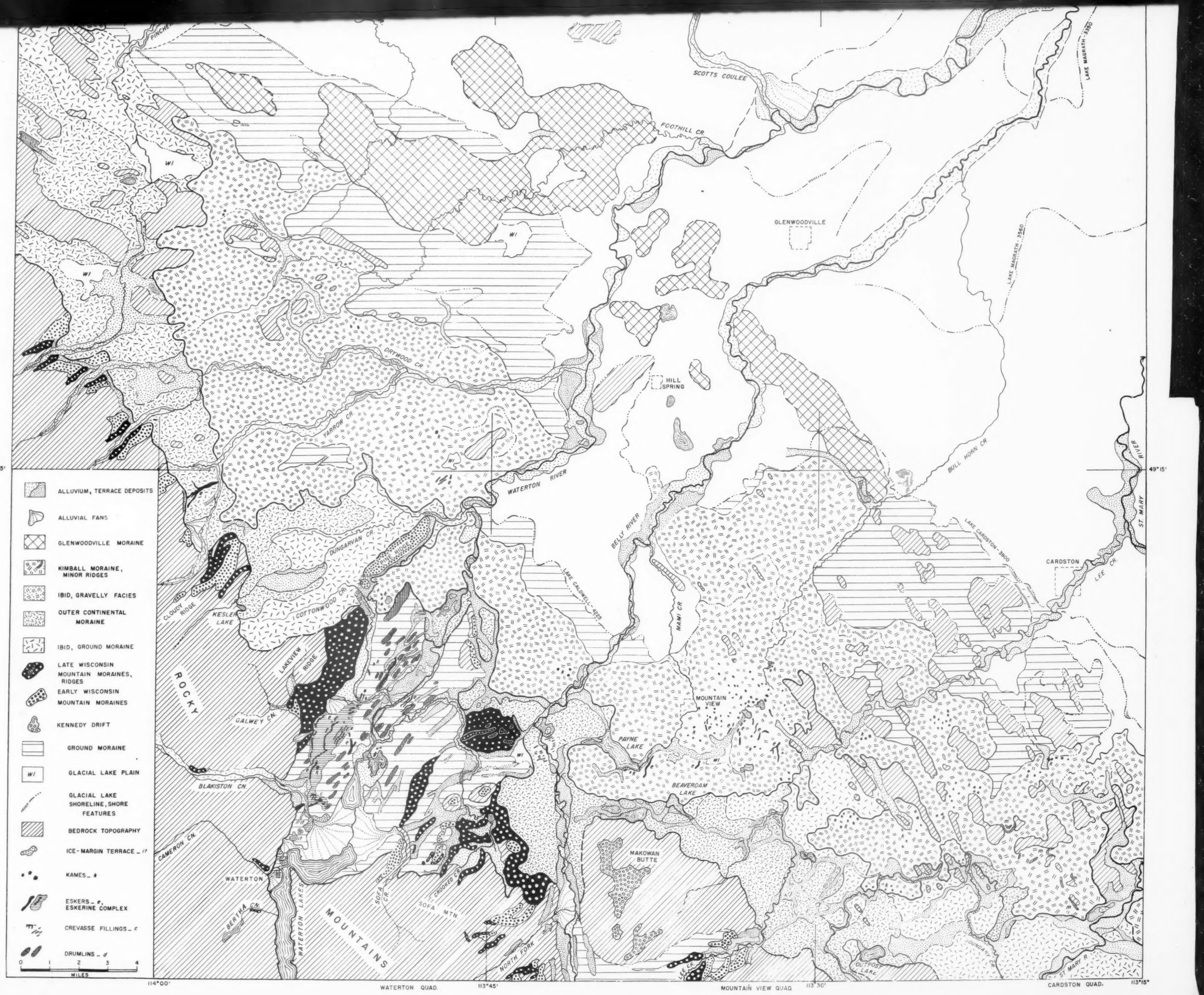


GLACIAL LAKES OF WATERTON REGION

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# ROCKY MOUNTAIN AND CONTINENTAL PLEISTOCENE DEPOSITS IN THE WATERTON REGION, ALBERTA, CANADA

BY LELAND HORBERG

## ABSTRACT

Continental drift of probable Cary age overlies Early Wisconsin and Late Wisconsin mountain drift in the Rocky Mountain piedmont of the Waterton region, Alberta, Canada. On the basis of geomorphic relations and soil profiles the Early Wisconsin moraines are correlated with the Iowan and Tazewell, and the Late Wisconsin moraines with the Cary. An older, deeply weathered drift (Kennedy drift) on remnants of the high-level Flaxville erosion surface is correlated with the Kansan. There are marked differences in the profiles of weathering developed on drifts correlated with the Kansan, Iowan-Tazewell, and Cary. A buried soil, named the Drywood soil, which occurs on Early Wisconsin mountain drift below the continental drift, is tentatively correlated with the Brady soil of the Middle West.

A series of proglacial lakes, formed as the continental ice withdrew down drainage, are reconstructed on the basis of shoreline features and outlet channels.

Periglacial features are represented by natural mounds and involuted zones.

Postglacial deposits include: (1) colluvium and a volcanic ash layer assigned to the Xerothermic period, and (2) cirque moraine, gravel fans and valley gravels, and a buried peat bed assigned to the Little Ice Age. The peat bed has a radiocarbon age of 3261 ± 250–3327 ± 320 years.

In general the Pleistocene record suggests that glacial maxima in the mountains were essentially synchronous with the Kansan, Iowan-Tazewell, and Cary maxima of the Keewatin ice sheet in the western part of the continental interior.

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## Introduction

### *Purpose and Scope*

The Waterton area is situated along the Rocky Mountain front just north of Glacier National Park in southwestern Alberta (Fig. 1), in a region where Rocky Mountain glacial deposits occur in association with drift of the continental ice sheet. The area is thus critical in terms of regional Pleistocene geology, and the aim of the present study is to clarify correlations of Pleistocene events in the Rocky Mountains with the standard North American chronology of the Mississippi Valley.

The area lies within the 49° and 49°30′ parallels and the 113°15′ and 114°7′ meridians and

includes about 1200 square miles. Airplane photos and the Waterton Lake, Mountain View, Cardston, Beaver Mines, Pincher Creek, and Glenwoodville 15-minute topographic sheets were used as a base in detailed mapping. In reconnaissance studies the area was extended to include most of the Lethbridge Sectional Map and adjoining areas in Montana. A previous study in the Lethbridge region (Horberg, 1952b) clarified the stratigraphy of the continental drift sheets and provided a valuable background to the problems in the Waterton area. About 15 weeks were spent in the field during the summers of 1950, 1951, and 1952.

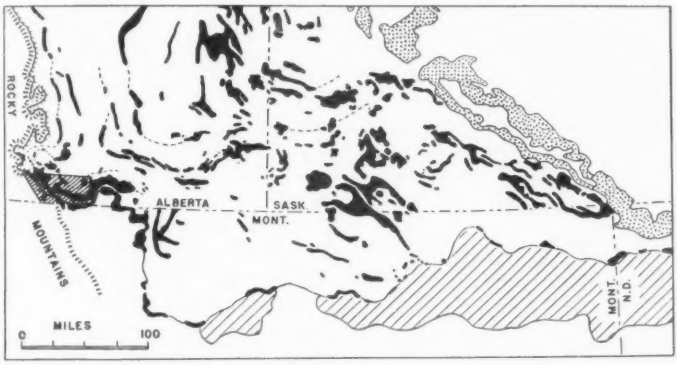



FIGURE 1.—REGIONAL GLACIAL MAP AND INDEX MAP

Iowan or Iowan and Tazewell (?) drift—diagonal lines; Cary moraines—black; limit of continental drift (Cary) in Canada—hachured line; Mankato Altamont moraine—stipple. Mountain drift not shown. Based on Bretz (1941) and Flint *et al.* (1945).

### *Acknowledgment*

Grants for field expenses from the Universty of Chicago in 1950 and 1951, and from the Penrose Bequest of The Geological Society of America in 1952 are gratefully acknowledged. The writer is also indebted to former students Warren Peterson and Charles N. Brown for assistance in the field; to Robert J. W. Douglas of the Geological Survey of Canada for conferences in the field and for reviewing the manuscript; to J Harlen Bretz for reviewing the manuscript; to James Thorp for critical reading of the sections on soils; and to L. R. Wilson for identifications of fossil wood and pollen.

### *Regional Setting*

The Waterton area can be divided structurally and geomorphically into three sections (Pl. 2): (1) the Lewis Range in which large-scale thrust faulting involves about 15,000 feet of Precambrian dolomites, limestones, argillites, and quartzites of the Belt series (Willis, 1902; Daly, 1912, p. 47–96; Hume, 1933); (2) a foothill zone of imbricate-fault structures paralleling the mountain front, characterized by discontinuous hogback and valley topography in Lower and Upper Cretaceous sandstones and shales (Allan, 1937; E. P. Williams, 1949); and (3) an area of relatively flat-lying shales and sandstones, largely Paleocene, which underlie the drift-covered plains and mesas in the northeastern part of the region.

The mountains reach a maximum elevation of 9600 feet and from their base elevations descend northeastward from about 5000 feet in the foothills to 3450 feet in the northeastern corner of the area. The area lies just north of the Hudson Bay-Missouri drainage divide, and the major streams—the Waterton, Belly, and St. Mary rivers—flow northeastward in the headwater area of the Saskatchewan River drainage basin.

In terms of regional Pleistocene geology the area includes the outer moraines at the southwestern margin of continental glaciation. These moraines are part of a moraine complex which is older than the Altamont moraine to the northeast and younger than the Iowan and Tazewell drift (Flint *et al.*, 1945) to the southeast in northern Montana (Fig. 1). Inasmuch as the

Altamont moraine is generally regarded as outer Mankato (Johnston and Wickendon, 1931, p. 36; Bretz, 1941, p. 34), the moraines occupy a position comparable to the inner Tazewell and outer Cary drifts of the Mississippi Valley.

The conditions of continental glaciation differed markedly from those in most of the central interior in that the ice advanced upslope into the Saskatchewan drainage basin and blocked the regional drainage and flow of meltwater away from the ice front. As a consequence a continuous series of proglacial lakes existed during the advance and retreat of the ice sheet. As the ice withdrew to the northeast, outlets at successively lower elevations along the ice margin and across sags into the Missouri drainage basin determined various lake stages. The remarkable coulees of the region—Lonely Valley, Verdigris Coulee, Etzikom Coulee and others—are abandoned valleys formed by the outlet streams (Pl. 6).

### *Climate, Soils, and Vegetation*

The climate varies considerably within the area, and this is reflected in the soils and vegetation. The average annual precipitation ranges from about 30 inches at Waterton Lakes at the edge of the mountains to about 15 inches in the northeastern part of the area. About half of the precipitation is in the form of snow. Temperature variations are extreme with warm summers and cold winters, the mean annual temperature varying from about 41°F. on the east to about 39°F. near the mountains (Wyatt *et al.*, 1939, p. 12–19).

Major soil groups are well developed, and in going from east to west one crosses the zones of Brown, Chestnut, and Chernozem soils of the plains (Wyatt *et al.*, 1939) to soils of the mountain complex dominated by Prairie and Brown Podzolic soils. Similarly from east to west there are vegetation changes from short-grass prairie, to mixed prairie and parkland, to mountain forests. Most of the area is open grassland, tree growth being largely restricted to the mountains and adjacent foothill slopes.

### *Previous Studies*

The early studies of Dawson and McConnell (Dawson, 1885; 1886; Dawson and McConnell, 1895) delineated the general position of the continental drift margin within the area and recognized the existence of former lakes. Also on the basis of regional observations in western Alberta they proposed the stratigraphic interpretation shown in Table 1, indicating that the continental drift is younger than the drift from the mountains. This fundamental relation was confirmed later by Calhoun (1906), Alden and Stebinger (1913), and Alden (1912; 1932, p. 67–72), although Dawson's correlations were not accepted (Table 1). In addition Alden and Stebinger recognized high-level remnants of pre-Wisconsin mountain drift older than any of the glacial deposits previously described.

In more recent studies Wickenden (undated map) mapped the glacial deposits of the eastern half of the Lethbridge map area, Allan (1939) briefly described some of the glacial features of the area from the point of view of soils, and Rutherford (1941) discussed the mountain derivation of large quartzite boulders on the plains of glacial Lake Cardston.

The bedrock geology and structure have been treated by Willis (1902), Daly (1912, p. 47–96), Hume (1933), Allan (1937), and E. R. Williams (1949). Current detailed studies of the area by R. J. W. Douglas of the Geological Survey of Canada are as yet unpublished.

## Geomorphology

### *Physiographic Divisions*

The major physiographic features of the area are determined by structure and pre-Wisconsin fluvial erosion. Upon these basic elements are superposed the landscape details of glacial deposition and sculpture. Thus three physiographic divisions—mountains, foothill ridges, and plains—which coincide with the structural divisions, may be recognized.

The Belt rocks of the mountains, although complexly faulted along warped, low-angle thrusts, have gentle dips for the most part and form gently sloping peaks and ridge crests determined by resistant stratigraphic units. Because of the varied lithology of the Belt formations, there is a repeated pattern of argillite slopes and valleys combined with ridges and cliffs of limestone, dolomite, and quartzite. There are no summit areas of subdued relief, and

all portions of the mountains in the Waterton area are in the early mature stage. The canyons, except as modified by glacial erosion, are youthful with steep walls and relatively narrow floors.

In the foot hills the topography is characterized by discontinuous hogbacks, isolated hills, and ridged uplands separated by broad basins and linear valleys (Pl. 2). In the Cardston quadrangle and in the northwestern part of the area there are subsequent valleys and ridges with marked northwest trends. Elsewhere structural control is not so marked, and in the west-central part of the area between Waterton River and upper Pincher Creek there are ridges of a second type with east-northeast trends controlled by pre-Wisconsin consequent drainage (Pl. 2). Although these ridges are drift-covered, bedrock is exposed in enough places to indicate that they are bedrock ridges rather than buried moraines of Rocky Mountain drift.

The plains in the northeastern part of the area are broad lowlands developed on weak, flat-lying sediments. Their present flatness, however, is due largely to a cover of glacial lake silts. The larger streams are sharply entrenched 150–200 feet below the surface of the plains.

### *Glacial Features*

The details of the landscape are the result of Wisconsin mountain and continental glaciation with erosion in the mountains, deposition in the foothills, and lake deposition on the plains.

In the mountains there is a full array of spectacular landforms resulting from glacial sculpture, but moraines are rarely encountered. For the most part the canyon floors are clean or partly covered by postglacial fan and alluvial deposits. Without doubt the most notable glacial feature of the mountains is Upper Waterton Lake, a remarkable glacial-trough lake over 400 feet deep (MacKay, 1940, p. 14) with a structurally determined threshold and narrows at its lower end (Waterton topographic sheet). Another striking feature compared with ranges farther south is the occurrence of cirques at low elevations well down on the mountain slopes. Many cirques floors lie at elevations between 5000 and 6000 feet and are only 1000 feet above the adjoining foothills.

The glacial landforms of the foothills and plains will be described in a later section and at this point it is sufficient to note that, like the Cary and Mankato drifts of the Upper Mississippi Valley, they are essentially unmodified by erosion and are characterized by numerous lakes and a lack of drainage integration.

### *Fluvial Erosion Surfaces*

Remnants of preglacial erosion surfaces were first recognized in the Lewis Range and adjoining piedmont by Willis (1902, p. 310, 344–46), who named and described the "Blackfoot peneplain" in Glacier Park from the highest benches near the mountains. In later, more-detailed studies by Alden (1932, p. 7) the term was dropped because of uncertainties in its designation, and the now widely-accepted chronology indicated below was proposed:

| | |
|---|---|
| No. 3 bench | Late Pleistocene |
| No. 2 bench | Early Pleistocene |
| Flaxville or No. 1 bench | Pliocene |
| Cypress Hills plain | Oligocene |

*Description.*—In Glacier Park the oldest surface, the Cypress Hills plain, is reported at only one place at an elevation of 8300 feet on Flattop Mountain near St. Mary Lakes (Alden, 1932, p. 7) (Fig. 2). There are no remnants of this surface in the adjoining piedmont, and the plains in this area have probably been eroded to below its projected level. The next lower Flaxville surface, however, is extensively preserved on the highest benches of the foothills at elevations of 6000–5000 feet and 1500–500 feet above present drainage lines.

In the Waterton region the Flaxville surface is represented by flat summit areas on Makowan Butte (Fig. 2) at elevations of 5700–6000 feet or 1200–1500 feet above Belly River. At this locality, as at many places in Glacier Park, the surface is mantled by pre-Wisconsin Kennedy drift. Other remnants of the surface along Belly River occur just south of the International Boundary and are represented by the crest of Lee Ridge and by spur remnants of rock benches along both the main valley and North Fork. Another possible remnant may be present on Lakeview Ridge north of Waterton Lake (Pl. 1) where the even crest of the east ridge truncates complex, imbricate structure in Belt rocks at about the Flaxville level (5600–5700 feet). No

flat summit area remains, however, and in absence of other remnants correlation is very uncertain

Farther east, flattish summit areas in the Milk River Hills, although largely drift-covered, have been correlated with the Flaxville plain (Alden, 1932, p. 14). They occur at elevations descending from 4600 to 4200 feet down drain-

TABLE 1.—PLEISTOCENE STRATIGRAPHIC INTERPRETATIONS

| Time | Waterton region | Lethbridge, Alberta Horberg (1952) | Montana-Alberta Alden (1932) | Southern Alberta Dawson and McConnell (1895) | Wyoming Blackwelder (1913) |
|---|---|---|---|---|---|
| Post-Xerothermic (Little Ice Age) | Alluvium, soil<br>Cirque moraine, gravel fans, valley gravels<br>Colluvium with humus layers, peat bed | | | | |
| Xerothermic | Colluvium, volcanic ash | | | | |
| Pre-Xerothermic | | | | | |
| Mankato | | | | | Pinedale |
| Two Creeks | | | | | |
| Cary | Periglacial features<br>Lake Lethbridge silts | | | | |
| | Lethbridge moraine | Upper till | | | |
| | Lenzie silts | Lenzie silt | | | |
| | Glenwoodville moraine<br>Lake silts<br>Kimball moraine<br>Outer Continental moraine | Lower till | Upper till | Upper boulder clay | |
| | Lake silts<br>Late Wisconsin mtn. drift | | | | |
| Brady | Drywood soil | | | | |
| Tazewell | Early Wisconsin mountain moraines | | Interglacial beds | | |
| | Saskatchewan gravels—local | | | | |
| Iowan | | Basal till | Lower till (?) | | Bull Lake |

TABLE 1.—*Continued*

| Time | Waterton region | | Lethbridge, Alberta Horberg (1952) | Montana-Alberta Alden (1932) | Southern Alberta Dawson and McConnell (1895) | Wyoming Blackwelder (1913) |
|---|---|---|---|---|---|---|
| Sangamon | | Saskatchewan gravels—other areas | | | | |
| Illinoian | | | | | | |
| Yarmouth | | | | Saskatchewan gravels | Interglacial beds | |
| Kansan | Kennedy mountain drift | | | Pre-Wisconsin mountain drift; Nebraskan or Kansan | Lower boulder clay | Buffalo |
| Aftonian | | | | | | |
| Nebraskan | | | | | Saskatchewan gravels = Albertan drift | |
| Preglacial | | | Saskatchewan gravels | | | |

age, and in the southwestern part of the hills are about 500 feet above Milk River. The surface and its relation to No. 2 bench are well shown in the unglaciated section of the hills north of North Fork Milk River. In this area a dissected but clearly defined, 200-foot scarp separates the Flaxville surface at about 4500 feet from No. 2 bench at about 4300 feet. Gravels composed of rounded pebbles and cobbles of Belt quartzite and argillite up to 8 inches in diameter with a few diorite and basalt are present on the Flaxville surface. The absence of limestone and dolomite, and the report of a rancher that a deep red soil 5–10 feet thick occurs on the upland flat are evidence of thorough weathering of the deposit.

The No. 2 bench of Alden is widely preserved at elevations of 5500–4200 feet along North Fork and Middle Fork of Milk River about 500 feet above drainage. From this area remnants of the surface can be traced northeastward into the southern part of the Milk River Hills where they become drift-covered. In the Waterton region the surface is best represented by flat ridge crests east of Makowan Butte, which lie about 800 feet below the Flaxville surface on the summit. The surfaces are 5000–5200 feet in elevation and about 500 feet above Lee Creek. West of Makowan Butte a corresponding surface, but with a drift cover, occurs in the headwater basin of Crooked Creek at an elevation of about 5000 feet and about 500 feet above Belly River (Fig. 3). Other possible remnants in the Waterton region include: spur remnants of old valley floors on the west side of Waterton Lake at elevations of about 5000 feet and 800 feet above the lake; Cloudy Ridge (Pl. 1) at elevations of 5200–5500 feet and 300–600 feet above Yarrow Creek; and, with less certainty, Pine Ridge and two related ridges (Pl. 1) to the northeast which rise 200–300 feet above surrounding plains. The three latter ridges are drift-covered, and the bedrock elevations are uncertain.

Alden's No. 3 bench is extensively developed along the North Fork Milk River at elevations of 4650–4450 feet and about 250 feet above stream level (Fig. 2). There is, however, only scant evidence of a corresponding surface in the Waterton region because of the cover of glacial drift. An additional complication arises from the fact that the Hudson Bay drainage in the Waterton region is about 1000 feet lower at the mountain front than the Milk River-Missouri drainage to the southeast. Hence, it is possible

that the No. 3 surface in the Waterton area was destroyed during a later cycle of erosion. The surfaces in the Waterton region which correspond most closely with No. 3 bench are the gressive denudation from the Pliocene on, rather than from an earlier time. Also the dating of the surface from vertebrate fossils at the type locality in the Cypress Hills is uncertain, inas-

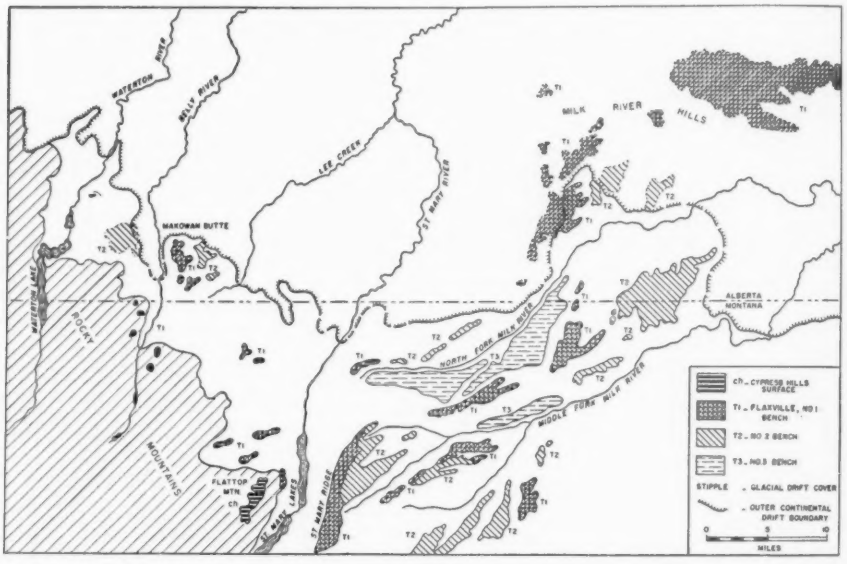



FIGURE 2.—EROSION SURFACES IN THE WATERTON-GLACIER PARK PIEDMONT
Mapping south of International Boundary based on Alden (1932, pl. 1)

broad, drift-covered basins between Dungarvan Creek and Drywood River and along Belly River north of Payne Lake (Pl. 2). The present streams are entrenched 150–200 feet below the adjoining uplands, and at many places bedrock occurs along the valley walls. Although in the field one is impressed by the uniformity of these surfaces, until subsurface data are available for the bedrock elevations the recognition of drift-covered erosion surfaces will remain uncertain.

*Age and origin.*—Alden (1932, p. 1–68) regarded the Cypress Plain as Oligocene, the Flaxville as Pliocene, the No. 2 bench as probably early Pleistocene, and the No. 3 bench as late Pleistocene. All the surfaces were considered to be the result of normal stream planation.

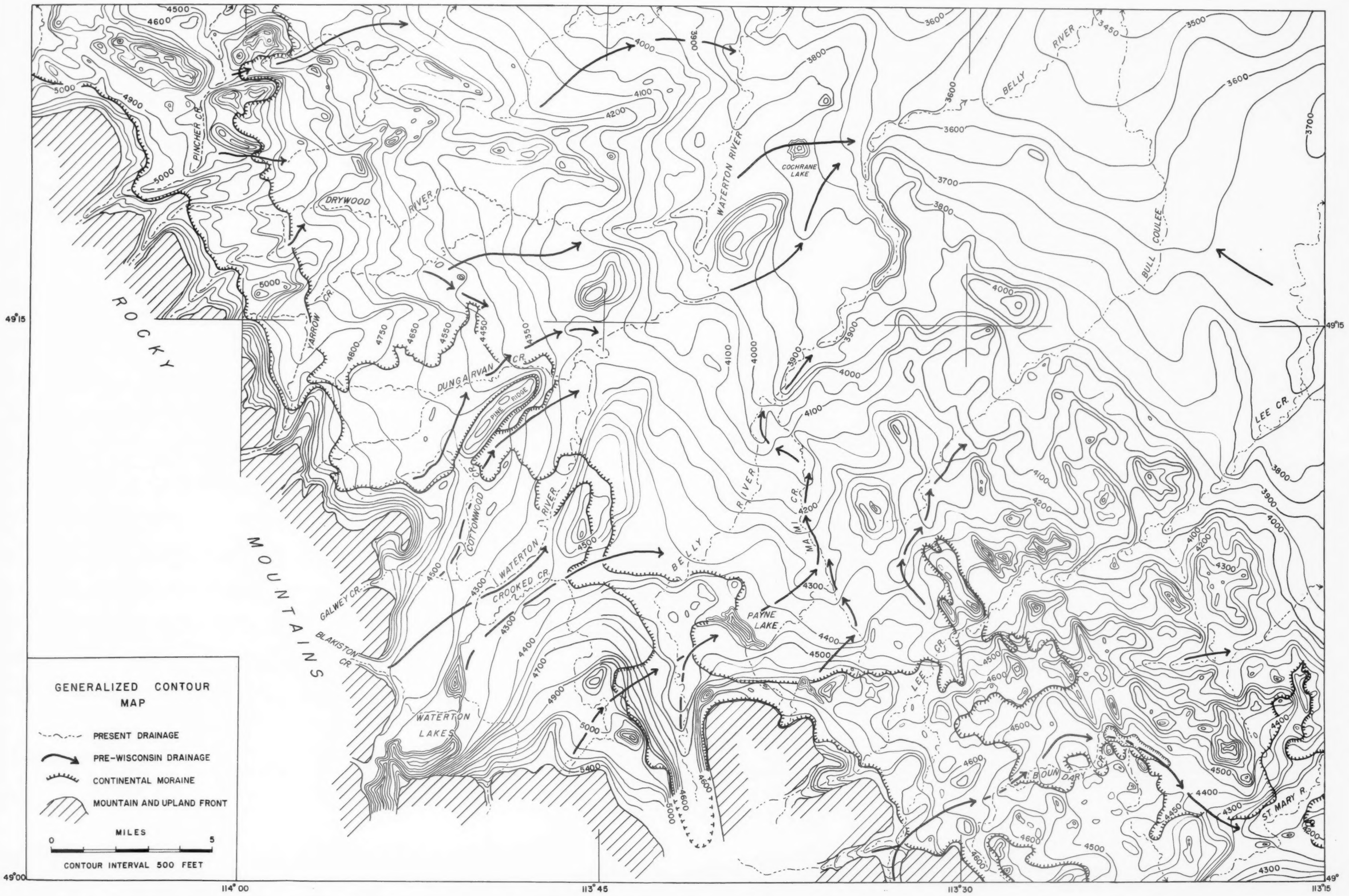
The present study provides little additional information. A possible Pliocene age for the Cypress Hills surface, however, is suggested by regional relations throughout the Rocky Mountain-Great Plains region which indicate pro- much as a younger gravel sheet, 40 feet thick, rests unconformably on the fossiliferous gravels assigned to the Oligocene (Mackin, 1937, p. 871–872). Hence, a younger surface probably truncates the Oligocene deposits.

The scanty evidence in the Waterton area suggests that both the No. 2 and No. 3 benches were eroded during the interval between the Kansan and Wisconsin glaciations.

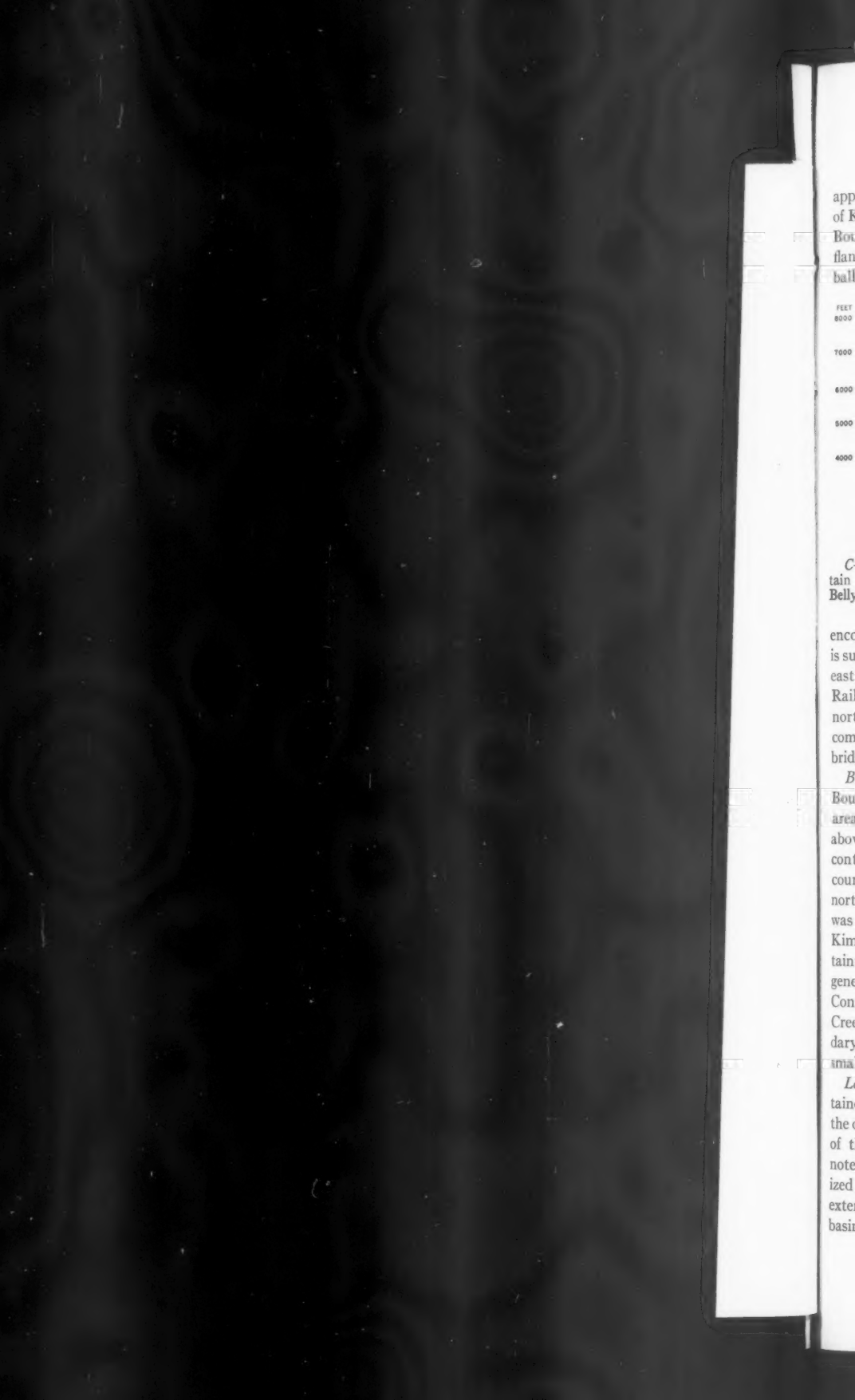
### *Drainage History*

The positions of preglacial valleys and drainage changes resulting from glaciation can be partially deciphered from the geomorphic evidence available. Most of the suggested major changes are indicated in Plate 2.

*St. Mary River.*—Although St. Mary River occupies a broad preglacial basin between the Milk River Hills and the foothill ridges south of Cardston, its present course within the basin




GENERALIZED CONTOUR MAP OF WATERTON PIEDMONT



appears to be largely postglacial. In the vicinity of Kimball, 4–6 miles north of the International Boundary, the river is clearly superposed on the flank of the upland to the west; north of Kimball the valley remains narrow, and bedrock is encountered at many places. For this reason it is suggested that the preglacial valley lies to the east along the sag occupied by the Alberta Railway and Irrigation Canal and continues north-northeast to Magrath where it becomes completely buried by glacial deposits (Lethbridge Sectional Map).

*Boundary Creek.*—The circuitous course of Boundary Creek in the southeastern part of the area (Pls. 1, 2, 3) and the narrows in bedrock above its mouth indicate derangement by the continental ice sheet. It is clear that the former course was the moraine-filled valley 3 miles northeast of the narrows and that the diversion was due to the advance of continental ice during Kimball glaciation. Although there is less certainty regarding the preglacial headwaters, the generalized contours and distribution of Outer Continental moraine suggest that upper Lee Creek valley formerly drained into the Boundary Creek basin. This is also indicated by the smaller size of Lee Creek valley.

*Lee Creek.*—Lee Creek appears to have maintained its preglacial course through the foothills, the only change being the postglacial acquisition of the headwater valley of Boundary Creek, noted above. Beyond the foothills the generalized contours suggest that the preglacial valley extended northward into the Bullhorn Coulee basin and to Belly River valley.

*Belly River.*—A major change along upper Belly River is indicated by the transverse course across the ridge shown by generalized contours west of Payne Lake (Pl. 2), and by the lobate projection of Kimball moraine into the old valley to the south. The valley through the ridge is narrow, and bedrock is exposed in many places. The preglacial valley probably extended northeast through the Payne Lake depression into the Mami Creek lowland and thence northward to present Belly River valley.

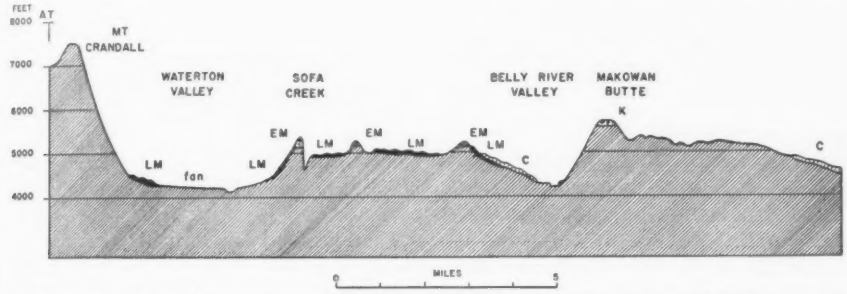



FIGURE 3.—CROSS SECTION SHOWING RELATIONS OF GLACIAL DEPOSITS TO EROSION SURFACES

*C*—Outer-Continental moraine; *LM*—Late Wisconsin mountain moraine; *EM*—Early Wisconsin mountain moraine; *K*—Kennedy drift on remnant of Flaxville surface; No. 2 bench between Sofa Creek and Belly River and east of Makowan Butte.

*Waterton River.*—A number of significant drainage changes in the upper Waterton valley are indicated by the anomalous courses of Cottonwood and Crooked creeks and by the narrows near the junction with Dungarvan Creek. The positions of preglacial valleys are suggested by bedrock narrows, positions of lowlands, and the lobation of the continental drift sheets (Pls. 1, 2). From these lines of evidence it is suggested that: (1) preglacial Waterton River followed a course along lower Crooked Creek valley northeastward into Belly River Valley; (2) Blakiston Creek, instead of turning south through the bedrock narrows, turned northeastward along the present eskerine depression and thence into the present valley of Waterton River; and (3) Galwey Creek turned northward along lower Cottonwood Creek valley and thence continued northeastward along the south side of Pine Ridge into present Waterton valley.

In the lower part of the valley the preglacial course probably led more directly eastward, south of Hill Spring ridge, into Belly River valley.

*Dungarvan Creek.*—Dungarvan Creek occupies the center of a preglacial lowland, but in its lower course it is superposed across the northeast end of Pine Ridge. Cottonwood Creek in preglacial time was undoubtedly a major tributary which followed a buried valley on the northwest side of Pine Ridge.

*Yarrow and Drywood creeks.*—In their upper courses these streams are in preglacial valleys. However, the lower course of preglacial Drywood Creek probably continued eastward through Cochrane Lake basin to Belly River. About 3 miles above the present mouth, Yarrow Creek is superposed on high bedrock, and a portion of the old, partially buried valley can be seen to the south. From this point it is uncertain whether the preglacial course swung back into Drywood Valley or continued southeastward into Dungarvan Valley.

*Pincher Creek.*—Pincher Creek follows a northerly course roughly parallel to the Kimball moraine and transverse to the east-west-trending, subsequent, preglacial valleys. These relations indicate that the present stream combines the headwaters of several preglacial streams, which formerly flowed eastward in valleys now blocked by moraine, and that diversion was caused largely by Kimball glaciation.

*Oldman River.*—Oldman River, north of the town of Pincher Creek, is superposed and entrenched on the south slope of the Porcupine Hills in a disproportionately small valley. However, a broad, structural valley, now blocked by Glenwoodville moraine and abandoned, extends southeastward toward Glenwoodville (Pl. 8). It is suggested, therefore, that the preglacial stream flowed southeast and was diverted to its present course by Glenwoodville glaciation.

## Areal Glacial Geology

### *Kennedy Drift*

*Description.*—The pre-Wisconsin mountain drift of Glacier Park (Pl. 3) was first described by Willis (1902, p. 328–330) as the Kennedy gravels and considered a fluvial deposit. Alden (1932, p. 32) later determined its glacial origin and included it in his pre-Wisconsin mountain drift. Inasmuch as the references are clearly to the same deposits, it seems appropriate that the original name be retained.

The Kennedy drift in the Waterton area is recognized only on Makowan Butte (Alden, 1932, p. 32–33) where it occurs on the Flaxville surface and at elevations some 200–300 feet lower (Fig. 3). Although significant erosion of the Flaxville surface occurred before deposition of the drift, this erosion probably did not extend as low as the level of No. 2 bench. Drift also occurs at much lower elevations as the result of extensive slumping on the underlying Cretaceous shales.

The drift, which is at least 100 feet thick, includes both till and outwash and in places is cemented by calcium carbonate into tillite and conglomerate. It is composed exclusively of Belt rocks from the mountains and includes boulders up to 6 feet in diameter. Although soils on the Kennedy drift will be discussed in a following section, it should be noted at this point that the upper 20 feet in places is weathered and leached, so that a residue of insoluble pebbles occur in a yellow-brown, limonitic, clayey matrix. Striated pebbles are not abundant, but striae can be found without difficulty on argillite pebbles. The deposit underlies the flattish summit area of the north butte, but even so it is extensively modified by deep weathering and slope processes so that no constructional forms remain.

*Interpretation.*—The geomorphic position of the drift close to the Flaxville level and its widespread although patchy distribution indicate that it is the remnant of a once-extensive deposit. Indeed it is possible, in view of the outwash character of ridge occurrences over much of the region, that there has been inversion of relief and that the present ridges have endured because of their gravel cappings. Alden (1932, p. 31–32) also found remnants of old drift on the No. 2 bench of the Glacier Park piedmont and suggested that either a single drift was deposited on both Flaxville and No. 2 benches or that two pre-Wisconsin drifts (Nebraskan and Kansan) were present. At present there is insufficient evidence to decide between these alternatives.

## Early Wisconsin Mountain Drift

### *General*

Moraines of mountain piedmont glaciers of Early Wisconsin age are represented by prominent lateral moraine ridges near the mouths of all major canyons from Lee Creek on the south to Pincher Creek on the north (Pl. 1). In addition possible remnants of the drift occur on bedrock hills east of Waterton Valley as much as 5 miles from the mountain front.

The moraines, where best developed, consist of composite ridges, which are widely breached so that the end moraines are eroded away and only the lateral ridges remain. The Late Wisconsin moraines, in contrast, are not so large and occur within or behind these moraine laterals at lower elevations. North of Waterton Valley the outer portions of the ridges are overlapped by Outer Continental drift, and their original extent is not known. Constructional forms are preserved on the ridges, but they are more subdued than those of the Late Wisconsin mountain and continental moraines. There is also more advanced dissection by minor drainage lines and better drainage integration than on the younger moraines. Soils on the Early Wisconsin mountain drift, which are discussed in a following section, confirm the geomorphic evidence and indicate distinctly that the deposits are older than the Late Wisconsin drift.

### *Description*

*Lee Creek.*—Composite lateral and breached end-moraine loops, which were deposited by Belly River ice that overrode the divide south of Makowan Butte, occur on upper Lee Creek near the International Boundary (Pl. 1). Three prominent Early Wisconsin moraine ridges with marginal channels occur on the south side of the valley, and a more complex series of at least five ridges is present on the north side. Distinct morainic forms are present but are subdued, and in many places the till is concealed by thick soils and colluvium. A valley-train terrace extends eastward from the outermost moraine on the north side of the valley.

A Late Wisconsin morainic loop lies just behind the Early Wisconsin moraines and closes off the valley. It consists of sharp knobs and basins with numerous ponds and is fronted by a valley-train terrace which extends through the Early Wisconsin moraines and valley train. These important relations are shown more clearly at this locality than at any other place in the region and are doubly significant because both terraces within a short distance down valley are overlapped and obliterated by the Outer Continental moraine. Stratigraphic evidence of the overlap is provided by exposures of continental drift overlying mountain outwash along a trail dugway on the east side of the valley at the north bend of the stream.

*Belly River.*—Early Wisconsin moraine remnants are present near the mouth of North Belly River, where they occur on both sides of the canyon about 1000 feet above drainage on possible remnants of No. 2 bench. There are two distinct lateral ridges on the north which can be related to combined ice streams in North Fork Belly River Canyon and a tributary canyon to the north. The inner moraine, which swings out toward the main valley, is an impressive ridge over 500 feet high. The moraines to the south represent lateral moraine remnants related to ice streams in North Fork Belly River canyon and in the main valley.

Youthful Late Wisconsin moraines occur at slightly lower elevations in the same areas and overlap the end of the main Early Wisconsin ridge on the north side of the canyon. The relations of the two sets of moraines indicate that the older moraines were trimmed back and extensively eroded by the younger valley glaciers.

*Sofa Creek and Upper Crooked Creek areas.*—Early Wisconsin mountain drift in this area is represented by lateral moraine remnants on both sides of Sofa Creek, and probably by the older drift capping two bedrock hills to the north and a third bedrock hill (Lookout Butte) on the east side of Crooked Creek basin. The west Sofa Creek moraine occurs on a rock bench close to the No. 2-bench level, but the other remnants occur on hills 200–300 feet above the No. 2 surface in upper Crooked Creek basin (Fig. 3). Late Wisconsin moraines from various sources occur at slightly lower elevations and more or less completely surround the remnants of older drift.

Subdued morainic forms are recognizable on the Sofa Creek laterals, but the drift on the bedrock hills is without topographic expression. Inasmuch as the elevations are below the level of Early Wisconsin mountain glaciation and soil profiles are comparable (Pl. 7) to those on undoubted Early Wisconsin moraines the drift capping the bedrock hills is considered Early Wisconsin rather than Kennedy. Although the west Sofa Creek moraine is typical Early Wisconsin, the other moraine remnants are not so clearly defined as in other parts of the region.

At this point mention should be made of a prominent ridge at the northeast corner of Sofa Mountain, which parallels the Late Wisconsin east lateral of Crooked Creek. This ridge has the geomorphic expression of an Early Wisconsin moraine, but it is composed exclusively of rubble of local derivation. The steep slopes on its east side are due to slumping, and the ridge itself is a remnant of an old debris accumulation. Soil profiles (Pl. 7) on the ridge indicate that it is Early Wisconsin.

*Waterton Valley.*—Early Wisconsin mountain moraines were apparently largely destroyed by later mountain glaciation, and only remnants of drift remain at the mouth of Galwey Creek, on Pine Ridge, and on a prominent hogback east of Waterton River (Pl. 1). The drift on the hogback has been reduced to a lag of pebbles and boulders on the bedrock surface. On Pine Ridge, however, there is a thick mantle of drift, but, except for low knolls and swales on the flat summit, the topography consists of smooth erosional slopes. Bedrock is exposed only at the northeast end of the ridge, so that the thickness of drift is uncertain. Both ridges are bordered at lower elevations by youthful continental moraines, which abut against them with marked geomorphic unconformity. Soil profiles on Pine Ridge drift are similar to profiles elsewhere on Early Wisconsin moraines.

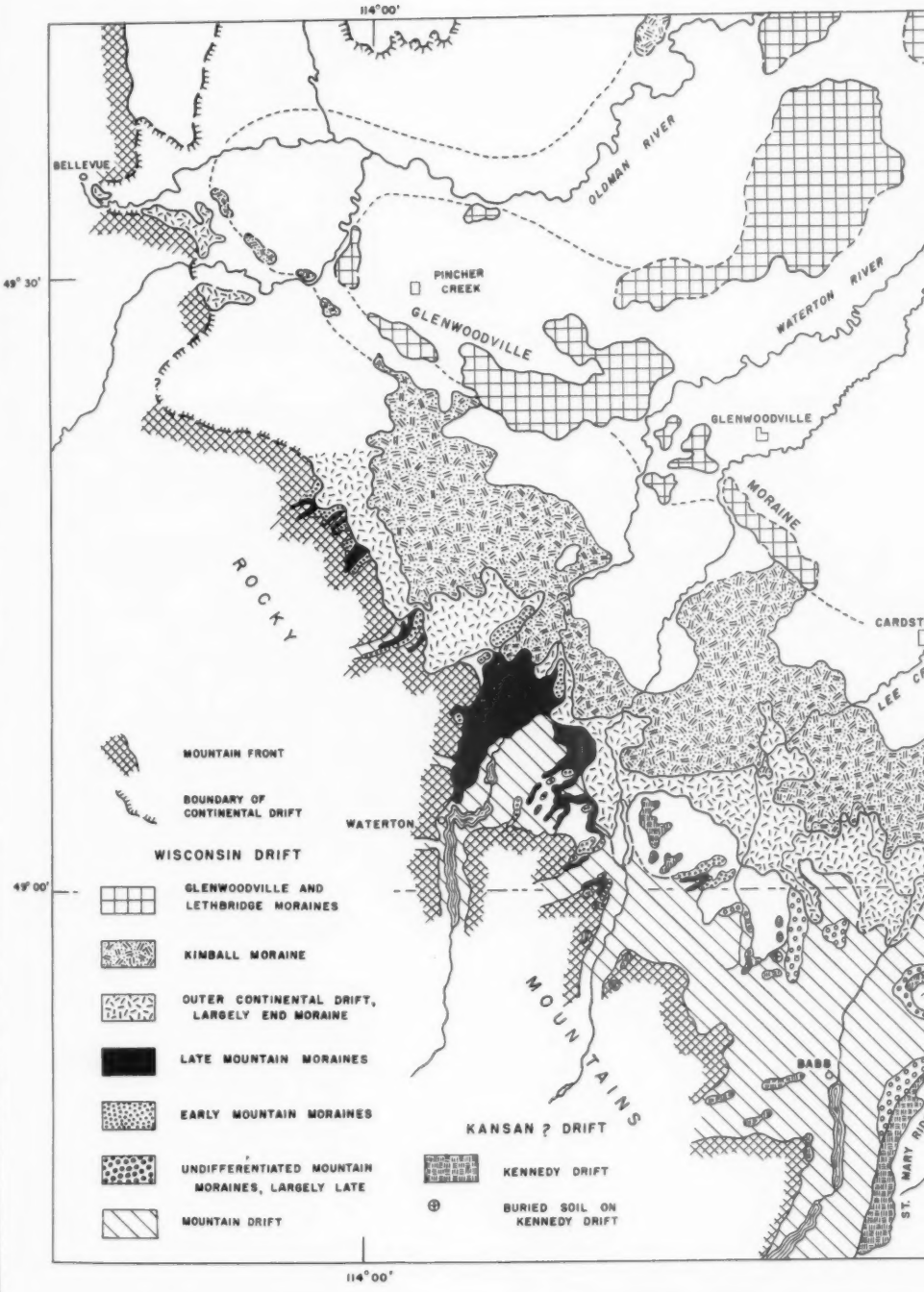
*Cloudy Ridge and Yarrow Creek.*—Although thick glacial deposits cover most of Cloudy Ridge, the relations to bedrock close to the mountain front indicate that it is fundamentally a bedrock ridge close to the level of No. 2 bench. Both Early and Late Wisconsin mountain moraines occur on the summit and west flank, and these in turn are overlapped by the Outer Continental moraine at the outer end of the ridge (Pl. 5, fig. 1).

The Early Wisconsin moraine consists of low eroded ridges with faint traces of knolls and swales, whereas the Late Wisconsin ridges are sharp and hummocky. The moraine was deposited largely by ice from South Yarrow Canyon, although ice from Dungarvan Canyon to the south may have contributed. The corresponding Early Wisconsin lateral along the mountain slope to the west occurs at a lower elevation and is less conspicuous. Farther north on North Yarrow Creek a large Early Wisconsin morainic ridge extends out from the mountain front on the south side of the valley.

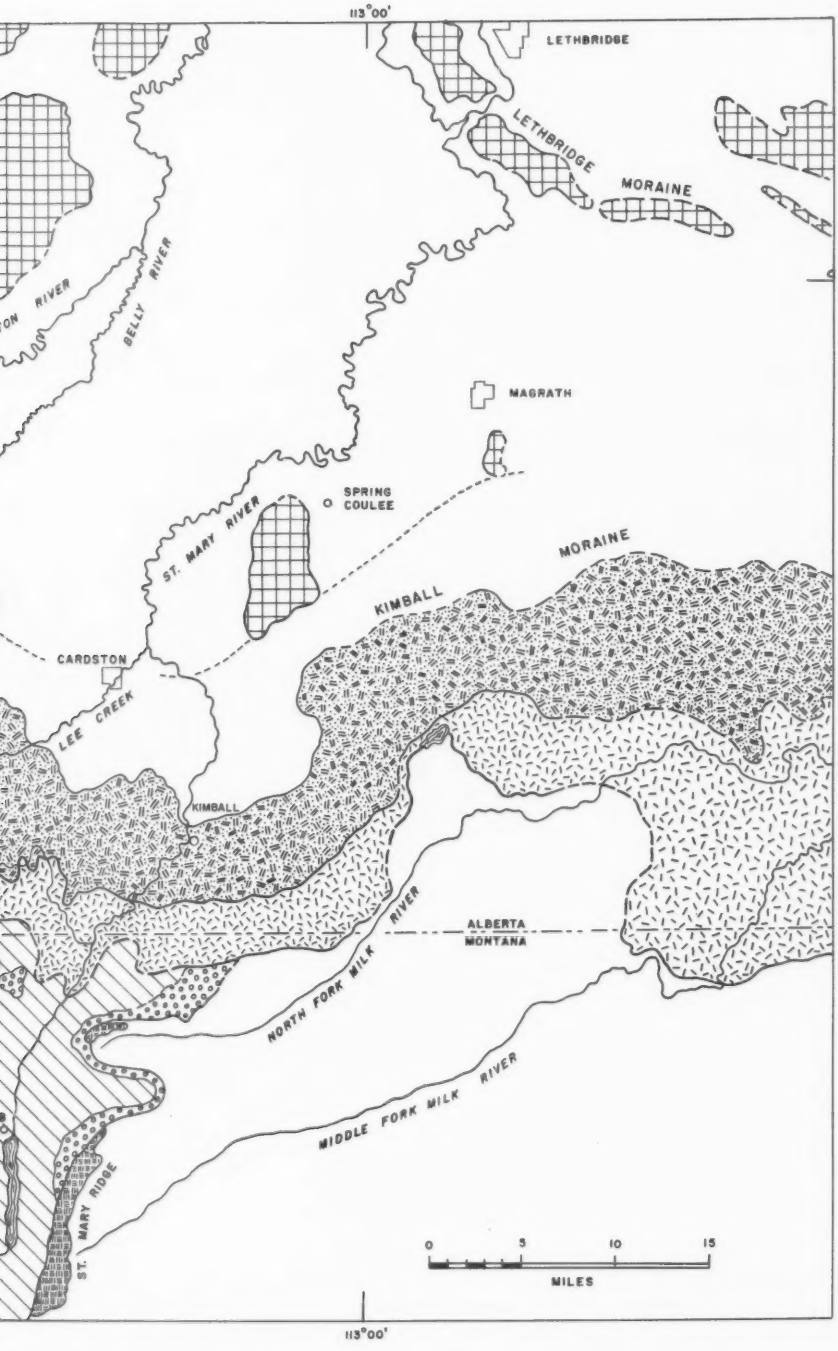
*Drywood and Pincher creeks.*—The relations of Early and Late Wisconsin mountain moraines and Outer Continental moraine at the mouths of North and South Drywood and Pincher canyons have the same consistent pattern as those farther south. For the most part the Early Wisconsin moraines are preserved as remnants of massive laterals on bedrock ridges which extend out from the mountain front, and it is difficult to determine the relative bulk of bedrock and drift. The slopes are eroded and subdued, and pronounced scarps, separating them from the Late Wisconsin moraines, record significant interstadial erosion. The Late Wisconsin lateral moraines occur within and below the older ridges, and Outer Continental moraine overlaps the extremities of both.

### *Character of Drift*

The Early Wisconsin mountain drift is composed almost exclusively of Belt constituents, varying somewhat in proportions from canyon to canyon. The till in most places is hard and stony and has a sandy matrix. Beds of sand and gravel are commonly associated with the till and in places form cemented ledges. A purple-gray color derived from red and purple argillites is characteristic. Stones on the surface are commonly fragmented by frost action. In places the limestones and dolomites show solution facetting and pitting, and the diorites have rinds of weathering up to an eighth of an inch thick. Below a depth of about 60 inches the till is uniformly calcareous.




GLACIAL DEPOSITS OF WATERTON–GLA
MAPPING SOUTH OF INTERNATIONAL BOUNDARY BASI




RTON–GLACIER PARK REGION
OUNDARY BASED ON ALDEN (1932, PL. 1)

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### *Interpretation*

The Early Wisconsin moraines are remnants of an extensive drift sheet deposited by piedmont glaciers that extended far out into the foothills. The present restricted surface distribution of the drift is due to the overlap of continental drift and, to a less extent, to interstadial stream erosion and to erosion by Late Wisconsin mountain glaciers.

The original extent of the drift may be inferred directly from exposures of mountain till below continental till, 6 miles from the mountain front on Drywood Creek, 7 miles on Dungarvan Creek, and 7 miles on Belly River (Pl. 8).

## Late Wisconsin Mountain Drift

### *General*

The Late Wisconsin mountain moraines and their relations to Early Wisconsin moraines and the continental drift have been described previously and, except for the moraine in Waterton and Belly River valleys, it will suffice to summarize their outstanding features. (1) They are more youthful than the older moraines and contain sharp morainic forms with numerous undrained basins. (2) In places, as in upper Crooked Creek basin and Lee Creek, the full morainic loops are essentially intact. (3) Surface stones are less fragmented and weathered than those on the older moraines, and leaching is not so deep, but the character of the fresh till is much the same. (4) They are smaller and occur below and within or behind the older mountain laterals. (5) They are overlapped by the Outer Continental moraine.

### *Belly River Moraine*

The extensive moraine on the west side of Belly River illustrates the sensitive marginal lobation of the ice as it emerged from the canyon into the foothills. The moraine belt first protrudes westward into a small basin to block out the headwater area of east Crooked Creek, thence loops back around a bedrock hill (Lookout Butte), and from there extends westward into Crooked Creek basin as a distinct morainic loop. To the north and east the moraine is covered by continental drift. The moraine is composed of hummocky moraine and of discontinuous ridges that tend to parallel the drift margin.

The relations of moraines in the upper Crooked Creek basin are of particular interest because of the crowding together of ice lobes from Belly River valley, from the cirque at the head of Crooked Creek, and possibly from Sofa Creek valley (Pl. 1). The Crooked Creek and Belly River moraine loops are clearly shown, and pebble counts (Fig. 5), as well as moraine outlines, indicate that the moraine in the central part of the basin was deposited by ice from Belly River valley rather than from Sofa Creek Valley. However, the existence of a basin open to the west north of the Belly River lobe in Crooked Creek basin, and a series of small lakes along its northern margin suggest that an ice lobe from Sofa Creek was also present. The lake basins occur above the level of the valley floor on the lower slope of a bedrock hill (Birdseye Butte) and are strikingly aligned and symmetrical. The southern rims of the basins are formed by low, even-crested embankments, and hummocky moraine occurs on the slopes above. Although the lake basins are atypical of moraine-dammed basins in some respects, it seems most logical to regard them as basins blocked on the south by push moraine from a Sofa Creek ice lobe. However, the alternative of blocking by Belly River moraine of an earlier stage than that represented by the moraine loop cannot be eliminated altogether.

### *Waterton Moraine Complex*

*Subdivisions.*—The drift complex in the Waterton valley may be subdivided into four types of glacial topography: (1) marginal areas of lateral and end moraine on the northwest and southeast sides of the valley; (2) a central area of ground moraine with drumlins and channels; (3) an interlobate eskerine tract which occupies a broad shallow depression eroded in the ground moraine; and (4) a large ice-block depression in the central part of the valley, which is bordered by ice-margin terraces and includes the basins of the lower lakes. The areal relations indicate that the features of these areas developed in chronologic sequence with initial

deposition of marginal moraine and ground moraine followed by subglacial erosion of the eskerine depression, esker deposition, and finally the formation of the ice-block depression. The superposition of esker, kame, and associated gravels on lodge till of the ground moraine supports the geomorphic evidence.

*Marginal moraines.*—The morainic area on the northwest side of the valley occurs on the slope of Lakeview Ridge and is characterized by discontinuous ridges paralleling the valley. East of the ridge, however, the moraine is hummocky and contains numerous kettles and small lake basins. At the mouth of Galwey canyon the ice encircled an Early Wisconsin lateral moraine remnant and, reversing it direction of flow, moved up-canyon for a short distance to deposit a low mass of moraine. To the northeast the moraine is overlapped by continental drift.

The marginal moraine on the southeast side of the valley includes an ill-defined elongate morainic tract on the valley slope between Sofa Creek and Crooked Creek and an extensive morainic area on flatter ground east of Crooked Creek. The morainic ridgings are clearly shown just east of Crooked Creek (Pl. 1) where they trend east-northeast, paralleling the valley and the direction of ice movement. In the outer portion some of the minor ridges swing to the north, suggesting an original terminal segment of the moraine. To the east the moraine is covered by Outer Continental drift.

The till of the moraines in most places is purple gray, calcareous, and stony, but east of Crooked Creek, because of Cretaceous bedrock to the west, it becomes less stony, silty, and gray in color and more closely resembles continental till.

*Ground moraine.*—The ground moraine includes drumlinized till in the central lower portion of the valley and thin drift on glacially eroded and channelled slopes on the southeast side of the valley. The area of thin drift, which lies east of Lower Waterton Lake between Crooked Creek and Sofa Creek, is dominantly ice-scoured bedrock with low rock-drumlin forms, a few till drumlins at lower elevations, and channels in bedrock—all paralleling the direction of ice movement. West of Waterton River the till is much thicker, and the drumlins reach maximum development.

The drumlins are low, elongate ridges without pronounced asymmetry and have a uniform northeast trend. They vary in height from 20 feet to over 150 feet; the largest, at the northern end of the area, is about a mile long and was massive enough to obstruct the continental ice as it moved into the valley. Although younger lake silts cover much of the area, there are sufficient exposures to indicate that the drumlins and ground moraine in general are composed of calcareous, stony, purple-gray till (Pl. 4, figs. 2, 3).

The channels in the ground moraine are shallow troughs in till and bedrock. In most places they are discontinuous, but some are fairly continuous. They occur from 600 feet above the valley on the southeast down essentially to the valley floor. Although some linear depressions are possibly due to glacial scour and subglacial erosion, their parallelism with valley slopes and occurrence at successively lower elevations evidence an origin by marginal streams as the ice melted down in the valley. They are significant because they indicate that there was free drainage down-valley and that locally at least there was no obstruction by continental ice at the time of Late Wisconsin mountain glaciation.

*Eskerine complex.*—Eskers and associated kames form a distinct eskerine tract west of Waterton River, but there are also scattered occurrences on the ground moraine to the east and north.

The main eskerine tract is between Blakiston and Galwey creeks in the western part of the valley and is localized in an interlobate position between ice streams from Blakiston Canyon and the main Waterton Canyon (Fig. 4). In this area a broad depression was eroded slightly below the level of the ground moraine, and the eskerine complex is largely confined to this depression. Somewhat similar compound eskers in channels in drumlinized ground moraine have been described by Armstrong and Tipper (1948, p. 293–299) in north-central British Columbia.

The major features of the eskerine complex are as follows:

(1) The larger, clearly defined ridges form a parallel series which trends northeast parallel

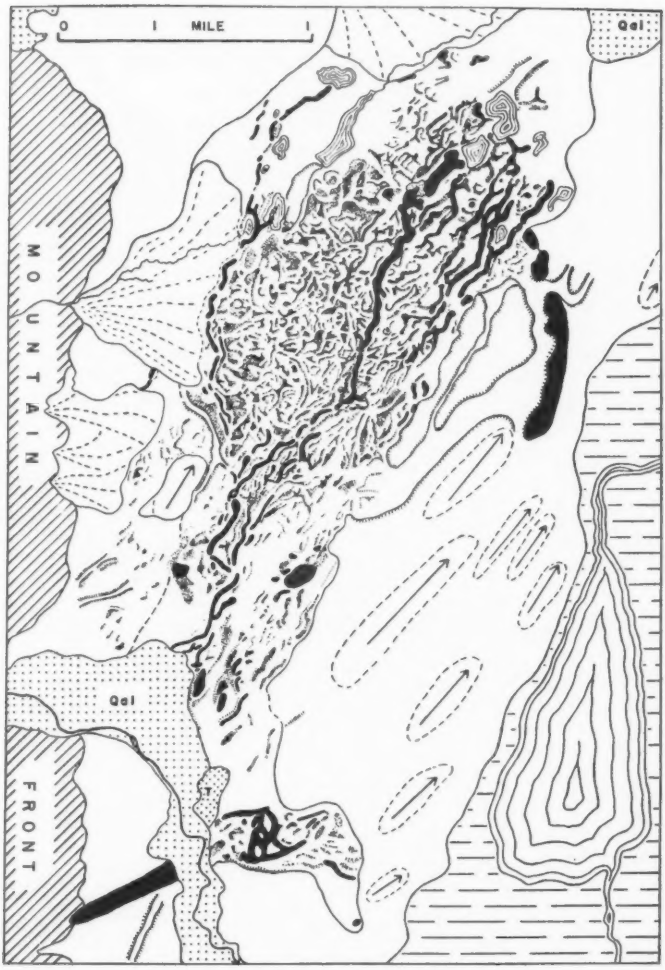


FIGURE 4.—ESKERINE COMPLEX IN THE WATERTON VALLEY

(Based on airplane photos.) Eskers, kames, and crevasse fillings—solid black; groundmat ridges in eskerine depression—stipple; drumlins—arrows; ice-block depression—broken horizontal lines; channels—hachured lines

to the direction of ice movement, as indicated by the drumlins.

(2) A maze of minor ridges in sinuous to angular patterns forms a ground mat on which the main ridges are superposed.

(3) Kames and ice-block depressions with small lakes are particularly prominent in the terminal portion (Pl. 5, fig. 2).

(4) The main ridges become more numerous and larger downstream, are commonly discontinuous, have both tributaries and distributaries, are bordered by discontinuous moats, and in places appear to be superposed on minor ridges.

(5) Channels, which in places lead into the eskerine depression, are marginal features.

(6) A prominent gravel ridge suggesting a large chute esker (Mannerfelt, 1945, p. 214–217)

descends the lower mountain slope south of the mouth of Blakiston canyon. It probably was formed by a subglacial stream which combined ice-margin drainage from both Blakiston canyon and the main valley.

(7) Crevasse fillings with an angular, network pattern are distinctive features at the upper end of the complex.

(8) The average local relief in the area is about 50 feet, and the maximum about 100 feet.

(9) The main esker is over 2 miles long, 30–200 feet wide, and is 60 feet high at its midpoint.

(10) In many places the ridges are distinctly segmented with mounds which tail out upstream to the next mound.

(11) The ridges are composed of purplish sandy to silty gravel.

The eskers on the ground moraine outside the complex are superposed on the drumlin topography and are smaller in size. The isolated small esker north of Galwey Creek, which is about 1 mile long but only 15–20 feet wide and 5–10 feet high, extends from a depression on the south up and over two drumlins and down into a linear channel on the north (Pl. 1). It is conspicuously segmented and boulder clad. Excavations revealed about 1 foot of bouldery rubble resting on pebble gravel with a silty sand matrix. Much the same relations are shown east of Waterton River by a group of three small eskers which likewise are superposed on drumlin topography. Here the ridges also are segmented and boulder clad and in places are bordered by shallow, elongate troughs.

Without entering into the details of the esker problem it may be suggested, in light of the above features, that most of the eskers were formed subglacially in stagnant or essentially stagnant ice and that the eskerine depression developed in an interlobate position which received marginal drainage from both Blakiston Canyon and the main valley. The channeling of the ground moraine, which preceded esker deposition, probably resulted from subglacial erosion at the periphery of active ice behind the stagnant terminal zone, as the boundary between the two zones retreated up the valley. A similar view was expressed by Andersen (1931) for "tunnel valleys" and eskers in Denmark. In line with this concept the ground mat of minor ridging may represent a transitional phase between channeling and deposition. The widespread, bouldery ablation moraine, and the high degree of preservation of the ridges are strong evidences of subglacial origin under hydrostatic conditions. However, the rectangular crevasse fillings at the upper end of the eskerine complex, as well as the isolated kames, are doubtless superglacial.

The discontinuous marginal troughs are probably relicts of segments of channels within which eskers were locally deposited. At any event this hypothesis is less difficult to apply in the Waterton area than in many other areas where there is not a clear-cut record of channeling that preceded deposition.

Another feature of uncertain origin is the beaded or segmented character of many ridges. Segmented eskers in other regions formed where the ice fronted bodies of standing water (summary in Flint, 1947, p. 153), but in this area all evidence points to free drainage away from the ice. Nevertheless, the persistence of segments in different eskers and their regular spacing strongly suggest that they are ice-margin features and record successive, possible annual, positions of the retreating ice front. Under conditions of free drainage the segments would represent alluvial cones deposited by small subglacial streams at the front of the ice where there was release from hydrostatic head and confining walls.

*Central ice-block depression.*—A large central depression, 100–200 feet below the level of the ground moraine, extends from the rock threshold at the mouth of Waterton Canyon northeastward to the mouth of Crooked Creek. The depression is about 5 miles long with a maxi-

PLATE 4.—ICE-BLOCK DEPRESSIONS AND GLACIAL TILL

FIGURE 1.—ICE-BLOCK DEPRESSIONS (*d*) AND KAME TERRACES (stipple) NEAR WATERTON
*f*—fan; *gm*—ground moraine; *c*—channel; *rk*—bedrock. Upper part of photo is to south. (Courtesy of Royal Canadian Air Force).

FIGURE 2.—MOUNTAIN TILL
Waterton ground moraine

FIGURE 3.—CONTINENTAL TILL
Outer Continental moraine in Belly River Valley.

FIGURE 1

FIGURE 2

FIGURE 3

ICE-BLOCK DEPRESSIONS AND KAME TERRACES NEAR WATERTON

FIGURE 3

FIGURE 2

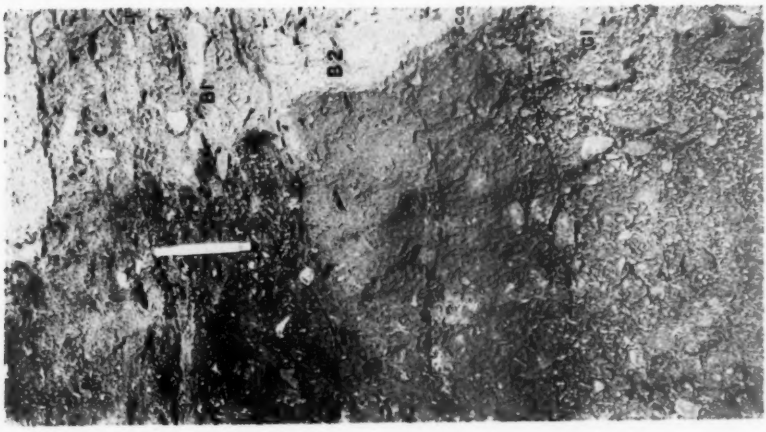


FIGURE 1

mum width of about 2 miles and contains the basins of Lower Waterton and Maskinonge lakes and the east arm of Upper Waterton Lake. It is partially filled by the large alluvial fans which determine the present outlines of the lakes. The western border of the basin is embayed by subsidiary ice-block depressions bordered by ice-margin terraces (Pl. 4, fig. 1), which provide clear evidence of a former stagnant ice margin. Altimeter elevations and heights of kame terraces above the lake from northeast to southwest are as follows:

| | | |
|---|---|---|
| Terraces east of Blakiston Creek | 4371 feet<br>4340 | 179 feet<br>140 |
| Terrace West of Blakiston Creek (golf course) | 4326 | 134 |
| Terrace S.W. of Lonesome Lake (gravel pit) | 4365 | 173 |
| Terrace N. of Waterton (Prince of Wales Hotel) | 4346 | 154 |

These elevations fail to show accordance and indicate that the terraces were separated by ice barriers and also, as in the case of the two terraces east of Blakiston Creek, that deposition occurred when the ice was at different levels.

The terraces are composed largely of coarse gravel with a sandy matrix, but there is a great range and abrupt changes in grade size. Boulders over 2 feet in diameter may occur with layers of sand, silt, and silty clay in a single exposure. The beds in most places dip toward adjoining depressions at angles of 30°–35°, and some are contorted and faulted. Along the terrace front east of Blakiston Creek and in the gravel pit near by the beds are cemented by calcium carbonate into conspicuous dipping ledges of conglomerate.

The depressions and related features record the waning phase of ice stagnation in the valley, and there are no comparable kame terraces in the canyon above. Final stagnation thus appears to have been localized in the center of the valley just below the rock threshold of the deep Lake Waterton glacial trough where conditions were particularly favorable for segmentation of the glacier into active retreating ice up valley above the threshold and a stagnant ice mass below.

### *Interpretation*

The Late Wisconsin mountain drift was undoubtedly less extensive than the Early Wisconsin drift, inasmuch as it occurs at lower elevations and in places end moraines are preserved. In the Waterton Valley it extended for at least 10 miles from the canyon mouth and in Belly River Valley at least 4 miles. Along the west margin of the Waterton Valley the upper limit of fresh, glaciated surfaces and moraine elevations indicate that in 7 miles the ice margin descended from elevations of about 5500 feet on Mt. Crandell near Waterton to elevations of about 4800 feet west of Pine Ridge—a gradient of about 100 feet per mile. With this projected gradient and no allowance for probable flattening, the Waterton glacier could have extended down valley beyond the east bend of the valley east of Pine Ridge, a distance of 15 miles from the canyon mouth.

The Late Wisconsin Waterton glacier must have been a compound glacier (Sharp, 1948) with individual ice streams from tributary canyons which maintained their identity down valley. Some evidence of this is provided by the position of the moraines, the linear depressions, and the eskerine complex. On the basis of these it is suggested that: (1) the main ice stream from Waterton Canyon followed the middle of the valley; (2) an ice stream from Blakiston Canyon was present west of the eskerine complex; (3) an ice stream from Cameron canyon followed a depression at the foot of the mountains north of Waterton townsite; and (4) an ice stream from Sofa Creek was directed to the east into upper Crooked Creek basin and the area south of the main marginal moraine. The available pebble counts (Fig. 5), which reveal

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PLATE 5.—MORAINES AND PERIGLACIAL INVOLUTIONS

FIGURE 1.—EARLY WISCONSIN MOUNTAIN LATERAL MORAINE (*M*) OVERLAPPED BY OUTER CONTINENTAL MORAINE (*C*) ON CLOUDY RIDGE

FIGURE 2.—ESKERINE COMPLEX IN LATE WISCONSIN WATERTON MORAINE

Crevasse fillings and ice-block depressions in foreground; eskers in background.

FIGURE 3.—INVOLUTED ZONE IN DEPOSITS OF GLACIAL LAKE MACLEOD

Masses of light-colored silt from above are enclosed by dark-colored clays. Scale indicated by 6-foot rule.

difference between Sofa Creek drift and drift in the main valley, are of interest mainly because they suggest how detailed studies of drift

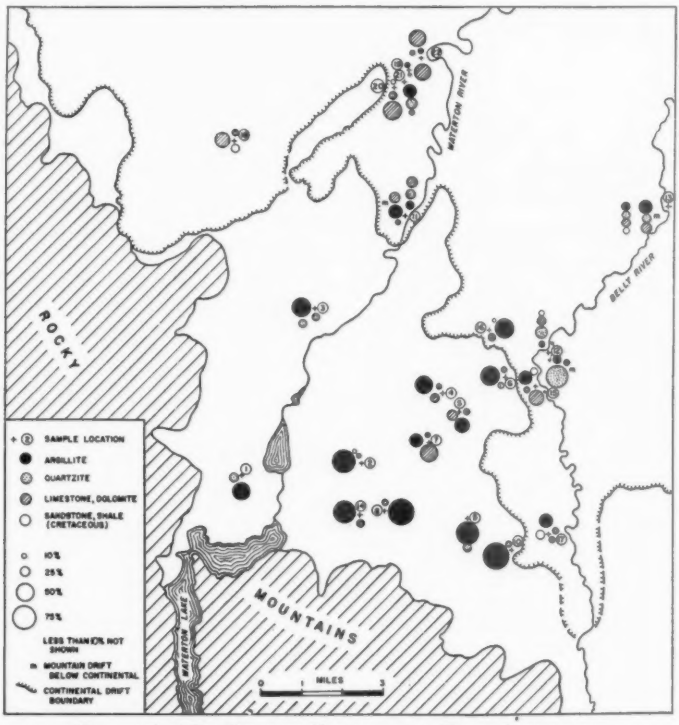



FIGURE 5.—MAP SHOWING VARIATIONS IN DRIFT LITHOLOGIES

Frequencies shown by proportional circles according to scale shown. Belt igneous, 13 per cent in sample 1, not shown.

defined end moraines were observed either by the writer or by R. J. W. Douglas (personal communication), who traversed the entire area in the course of areal bedrock studies. Scattered remnants of moraine, however, are present at the following localities (Waterton Lakes Park Topographic Map): Blakiston Canyon, about 3 miles above its mouth and at the junction with South Fork; Cameron Canyon, near its mouth and just below Cameron Lake; Bertha Canyon at about its midpoint; and Waterton Valley immediately above Waterton Lake. In addition there is a well-defined cirque moraine within the large cirque east of Mt. Blakiston.

Only the moraine in upper Blakiston Canyon was studied in detail. Here the moraine occurs as forest-covered lateral ridges and hummocky moraine for almost 2 miles along the canyon and can be related to an ice stream in the North Fork. The South Fork at its mouth is diverted

lithologies might serve as a test for the hypothesis.

The geomorphic aspect and soils of the Late Wisconsin moraines and continental moraines indicate that they are essentially of the same age. However, the available evidence from meltwater channels and valley trains shows that there was free drainage at the time of mountain glaciation. Continental ice may have been present contemporaneously at no great distance to the northeast, but there was a definite lag in its appearance within the local region.

## POSSIBLE YOUNGER CANYON MORAINES

The canyons within the mountains are relatively free of morainic material and no clearly

eastward by the lateral on the south side of the canyon. There is no evidence of an end moraine, and the low terrace which occupies most of the valley passes through the moraine remnants. Alluvial fans are superposed on the low terrace and record the last important episode of deposition in the canyon. A well-defined podzol soil profile (No. 25, Pl. 7; Soil profile 25, Appendix) with a leached zone of 25 inches is developed on the moraine. This compares closely with the depth of leaching on the Late Wisconsin mountain moraines and continental moraines in the piedmont, which, are considered Cary.

Later detailed studies may lead to recognition of younger stadial moraines in the canyons, but there is no evidence to support recognition at the present time. It appears that Mankato glaciation was very weak or locally absent and that a major discontinuity separates the Cary deposits from the cirque moraine and valley and fan gravels attributed to the Little Ice Age.

## Outer Continental Drift

### *Continental Drift Boundary*

The outer boundary of the continental drift in the region was first indicated on a map by Dawson (1885, Accompanying Geological Map of the Region in the Vicinity of the Bow and Belly Rivers), which showed the "approximate western limit of Laurentian and Huronion boulders". This boundary, although generalized, closely approximates the boundary shown in Plate 1 in the western part of the area. East of Waterton Valley the boundary was not indicated. In later studies Calhoun (1906, Pl. 1) showed a generalized boundary east of Waterton at a position somewhat north of the boundary recognized in the present study, and Alden (1932, Pl. 1) mapped the drift margin in detail in the area just south of the International Boundary.

The margin of the continental drift in most places is readily recognized by the terminal moraine, the clayey, nonstony character of the till as compared with mountain drift (Pl. 4), the presence of crystalline erratics, ice-margin channels, and moraine-dammed lakes in preglacial marginal valleys (Kesler and Outpost lakes, Pl. 1). The criterion of crystalline erratics, however, should be applied with caution because berg-carried erratics are widely dispersed in proglacial lake basins in front of the drift boundary.

The outer continental drift boundary in the eastern part of the region crosses the International Boundary about 7 miles west of Middle Fork Milk River (Pl. 3) and continues north along the edge of a remnant of No. 2 bench past Lake Shanks to North Fork Milk River near the crossing of the Del Bonita-Magrath road (Lethbridge Sectional Map). Beyond North Fork it loops to the north past Ross Lake, forming a re-entrant around a remnant of the Flaxville surface, before continuing south to recross the International Boundary. In this area continental till is exposed at the west edge of North Fork Milk River valley at the crossing of the Del Bonita-Kimball road, but the bulk of the moraine lies on the west side of Hudson Divide Ridge to the west. It appears that the margin crosses to the northwest side of Hudson Divide in the vicinity of the International Boundary. Thence it continues west along the north side of the divide, as mapped by Alden, into St. Mary Valley (Browning quadrangle). Recent excavations in continental till along the Babb-Cardston road indicate that lobation into St. Mary Valley was more pronounced than previously recognized, and the boundary is therefore shifted about 3 miles to the south. On the west side of the valley the margin crosses into Canada about 2 miles west of lower Boundary Creek. It then loops to the north around foothills and past Outpost Lake, but re-enters the United States as a narrow lobe which extends southeastward into the upper Boundary Creek-Pike Lake valley to a point 1½ miles southeast of Pike Lake (Pls. 1, 3; Cardston and Browning quadrangles).

West of upper Boundary Creek the drift margin continues westward along the base of a conspicuous foothill scarp into Lee Creek valley, which it crosses at the west bend of the valley (Pl. 2; Mountain View quadrangle). From this point on the drift margin is delineated by more detailed mapping (Pl. 1), and only the outstanding features will be noted. West of Lee Creek the boundary follows the base of the eroded scarp below No. 2 bench on the Makowan Butte upland to the pronounced southward lobation into Belly River Valley. Drift in the valley bottom was removed by subsequent

erosion so that the southern limits of the lobe are uncertain. However, projected elevations and proglacial terraces indicate that it reached at least as far south as the junction of North Fork, which is about 2 miles north of the International Boundary. Earlier references by Calhoun (1906, p. 48) and Alden (1932, p. 105) to the continental terminal moraine in Belly River Valley "10 miles" north of the International Boundary doubtless refer to the prominent Kimball moraine front to the north.

Beyond Belly River valley the drift margin recedes to the north away from the mountains around the hogback east of Waterton River. Because of obstruction by bedrock hills only a minor lobe extended into upper Waterton Valley. North of Waterton Valley the drift margin again extends to the mountain front and continues northwestward at the base of the mountains.

The boundary was not mapped in any detail beyond Pincher Creek, but reconnaissance observations indicate that it extends west into a structural lowland at the foot of the mountains south of Castle River (Pl. 3; Beaver Mines quadrangle) and that thence it continues north, bordered by discontinuous moraine, into Crowsnest Valley where a lobe extended to within 1 mile of Bellevue (Cowley and Blairmore quadrangles). North of Crowsnest Valley the margin diverges from the main mountain front, loops northward to form a large lobe in upper Oldman Valley (Douglas, 1950, p. 6–7), and thence swings south around the Porcupine Hills to the east.

*Elevations along the Continental Drift Margin*

Elevations along the margin of the continental drift can be determined from the generalized contours in Plate 2. In general the elevations within the region rise from about 3900 feet near Cut Bank, Montana, to over 5000 feet between Waterton and Pincher Creek, and thence descend to 4200–4400 feet to the north. However, there are important local variations in elevation which suggest the following alternative explanations: (1) obstruction by Late Wisconsin mountain ice, (2) deformation, and (3) obstruction and differential resistance to ice flow determined by the preglacial topography. As previously noted, there is no evidence to support (1), and (2) can be eliminated for local areas at least by the unsystematic variations in elevations and by the fact that the glacial lake shore lines of the region show no evidence of deformation. Turning then to topographic control as an explanation the following interpretations are proposed: (1) lower elevations (4150–4500 feet) in the upper Oldman drainage reflect the narrowness of the valley, the fact that it diverges from the major regional direction of ice movement, and the obstruction provided by the Porcupine Hills and the upland between Waterton and Oldman rivers southwest of Macleod; (2) maximum elevations (5000–5200 feet) to the south between Pincher Creek and Dungarvan Creek (Pl. 2) are due to the presence of broad avenues for regional ice flow provided by the Belly River and Waterton River lowland in the northeast part of the area and the absence of prominent foothills in front of the mountains; (3) lower elevations (4200–4500 feet) in the upper Waterton Valley clearly reflect the obstruction offered by the bedrock hills on either side of the valley and possibly also the bedrock hills to the northeast; (4) higher elevations (4900–5000 feet) in upper Belly River valley are determined by the broad open lowland along the lower valley; local lower elevations (4500) west of Payne Lake are due to the buried bedrock ridge immediately to the north; (5) generally lower elevations (4800–4600 feet) between Belly River and St. Mary River are explained by the wide foothill belt to the northeast; (6) still lower elevations (4400–4500) in St. Mary Valley reflect the divergence in trend between the valley and the direction of ice movement and deflection of the ice by the Milk River Hills; (7) progressively lower elevations (4600–3900 feet) with local variation to the east and southeast as far as Cut Bank, Montana, may be attributed largely to shadow effects from the Milk River Hills and possibly the Sweet Grass Hills to the northeast.

These suggestions are only of broad significance, and more remote topographic influences and other variables, such as mutual interferences of ice streams and shifts in directions of ice movements, should also be considered.

However, the relations appear to justify the significant general conclusion that postglacial uplift and deformation are not required to explain the variation in elevations.

### *Description*

The topography of the Outer Continental drift is similar to the Late Wisconsin mountain drift and the other continental drift sheets in terms of postglacial modifications, and the contrasts that exist are due to differences in initial forms. Within the area mapped in detail, the drift border is marked by youthful end moraine except in two areas, one south of Cottonwood Creek in the western part of the area and the other near Outpost Lake in the southeastern part of the area. In most places the moraine is not so abrupt and knobby as the outer Kimball moraine, but this is due in large part to its nongravelly character. Low mounds and ridges, swales, lakes and swamps, and poor drainage are characteristic. Gravelly facies are generally absent but do occur in the southern part of the Belly River lobe and elsewhere in restricted localities.

Crevasse fillings in the form of low ridges, which are strikingly parallel and close set, occur at one locality west of Payne Lake (Pl. 1). They were noted first on airplane photos and probably would not have been recognized from the ground. The ridges trend uniformly N. 60° E., are about 200–1500 feet long, 15–30 feet wide, 2–10 feet high, and have 3- to 4-foot boulders on their crests. They are situated on the south slope of a bedrock ridge (Pl. 2) where crevassing could logically be anticipated. Somewhat similar features in Quebec and Saskatchewan have been described as crevasse fillings by Mawdsley (1936) and Sproule (1939), although Norman (1938) regarded the Quebec ridges as annual moraines.

A single prominent esker north of Cloudy Ridge descends the mountain slope between 5000 and 4700 feet in a direction almost normal to the drift margin. Tributaries at the upper end, and the increase in size downslope indicate that it carried drainage eastward, opposed to the direction of ice flow, into Yarrow Valley and thence across the col at the north end of Cloudy Ridge.

The Outer Continental till is silty to clayey, calcareous, dark gray where unoxidized, and gray buff where oxidized. It is less stony than the mountain till and contains fewer crystalline erratics than the younger continental tills (Pl. 4, fig. 3). Most pebbles and boulders are Belt rocks from the underlying mountain drift, whereas the finer matrix appears to be derived largely from the local Cretaceous shales and sandstones.

### *Kimball Moraine*

*General description.*—The Kimball moraine is here named from the town of Kimball in the St. Mary Valley south of Cardston where typical end-moraine topography is present near town and in the area to the east. It is a well-defined unit throughout the area and in most places it is marked by an impressive moraine front. The front of the moraine closely parallels the Outer Continental drift margin and in most places lies less than 5 miles behind it. There are conspicuous minor lobations in the southern part of the Cardston quadrangle (Pl. 1), in Payne Lake preglacial valley, and in Waterton Valley where the Outer Continental moraine is locally overlapped completely. Elevations along the moraine margin (Pl. 2) are about 300 feet lower than the Outer Continental drift border but show parallel local variations.

The width of the moraine belt is variable, but in most places it is about 5 miles (Pl. 1). South of Hillspring, however, there are major expansions in width and pronounced offsets of the inner boundary. These relations suggest that the moraine in the ancient Belly River Valley is composite and that the expanded width is due to a local recessional phase which merges with the main moraine to the northwest and southeast. An earlier recessional phase is suggested by the variation in moraine width east of Payne Lake where a former stationary ice front probably margined the prominent meltwater channel east of the lake.

The topography of the Kimball end moraine is strongly suggestive of the Cary and Mankato moraines of the Great Lakes region. Initial ridges, knobs, and depressions are essentially intact, extensive surfaces are boulder strewn, drainage integration is minimal, and there are

hundreds of small lakes and ponds. In the central part of the moraine, as shown by the Mountain View quadrangle map (1 inch to 1 mile), 535 small lakes and ponds are present in an area of 22 square miles, which is almost 25 per square mile. The lake basins are largely due to irregular deposition of till or to melting of ice blocks in till or outwash with no relation to preglacial valleys. Other types are represented by Payne Lake, which occupies an ice-block basin in till localized by a preglacial valley, and Beaverdam Lake which is a moraine-dam lake (Pl. 1).

The ground-moraine topography is gently undulating with low swells and swales and, even in areas not mapped as lake plains, it is extensively covered by lake silts. The boundary between areas of ground moraine and lake plain commonly coincides with a major glacial-lake shore line, but in places there is a gradual transition from one type to the other.

Field observations indicate that the Kimball till is similar to the Outer Continental till except for a higher content of crystalline cobbles and boulders. Several counts of surface cobbles and boulders in the area between the Waterton and Belly rivers give an average of about 10 per cent by number for the Outer Continental till and almost 60 per cent for the Kimball till.

*Minor drift lobes in Cardston Foothills.*—In the southern part of the Cardston quadrangle the Kimball ice was obstructed and diverted by the higher ridges, so that a complex of minor ice streams controlled by structural valleys developed at the margin of the ice sheet. The most prominent lobes are indicated by the irregular moraine boundary (Pl. 1) and include: (1) the forked lobe which crosses Lee Creek from the northwest 5 miles southeast of Mountain View; (2) the large westward-protruding lobe a mile to the south; and (3) numerous minor lobes along Boundary Creek.

The lobes and associated features are particularly well shown in the Boundary Creek area where ice crossing the high, ridged upland to the north descended to the floor of the Boundary Creek basin as distinct "piedmont" lobes. Within the upland ice lobation is evidenced by small obstruction moraine ridges, which curve around the stoss ends of bedrock hills, and by minor ridges which parallel the margins of the lobes. Some larger ridges are indicated on Plate 1, but the details are best shown by airplane photos. Associated with the obstruction moraines are minor recessional moraines, which typically occur on the stoss slopes of cols between ridge crests. A composite moraine of this type is shown in Plate 1. These moraines were formed as the ice receded and the front became blocked by adjoining ridges.

The lobes on the valley floor are characterized by marginal ridges, kame moraine, and lobate gravelly facies. The gravelly deposits overlie till and are variable in composition, including large bodies of sand and some silt as well as gravel. In most places the topography is expressed as low knolls and swales with numerous ponds. Both marginal and intralobate meltwater channels extend back through gravel facies to till facies of the lobes. These features and the lobate outlines indicate that the deposits are not normal outwash deposits, but instead probably represent ablation moraine deposited on stagnant outer portions of the lobes.

The minor lobes and related features show that the marginal ice in this area was so thin that preglacial topography became a controlling factor and the higher ridge crests were not ice-covered. A short distance to the north, however, these features are absent, and ice undoubtedly covered all the highest summits.

*Kames and eskers.*— Kames and eskers are much more numerous in the Kimball moraine than in any of the other continental moraines and are particularly abundant in the Mountain View and Cardston quadrangles (Pl. 1). Most of the kames occur as areas of kame moraine along preglacial valleys, such as the preglacial Belly River Valley near Mountain View and the larger structural valleys in the Cardston quadrangle. However, large, isolated kames of moulin type are also present, as in the area west of Belly River.

Eskers commonly occur in association with the kames and are most numerous in the Mountain View area. In this area they are of particular interest, because they indicate that subglacial drainage was northeastward away from the ice front and eventually led into Lee Creek Valley. By this interpretation the meltwater channels, which enter Payne Lake from the

north and west, may be logically explained by a subglacial outlet into Lee Creek Valley. As a corollary it seems probable that the main channel, which leaves Payne Lake and ends blindly to the east, was initially a subglacial channel.

### *Glenwoodville Moraine*

The Glenwoodville moraine is here named from the town of Glenwoodville in the north-central part of the area. The moraine lies north of the foothills in the lowlands covered by glacial lakes Caldwell and Cardston and is represented by discontinuous morainic areas which have been modified by lacustrine deposition and erosion. In many areas the moraine is so extensively altered it would not be recognized as such except for relicts of morainic depressions and low knolls. The major processes involved in modification include: partial and complete burial of morainic forms by lake deposits; erosion of wave-cut scarps along shore lines; and truncation of morainic knobs and ridges by wave action.

As one would anticipate, the moraine varies greatly in topographic expression, and it is possible to differentiate areas of: (1) slightly modified moraine, (2) subdued moraine, and (3) relict moraine. In most areas of subdued and relict moraine the deeper basins still retain lakes, and there is commonly a greater concentration of boulders at the surface. From a strict physiographic viewpoint it is doubtful whether most of the Glenwoodville would qualify as "moraine", although its designation is fully justified in terms of historical interpretation.

Outside the area mapped in detail, the Glenwoodville moraine is represented by extensive morainic areas between Waterton and Oldman rivers and probably by isolated patches of moraine southwest of Spring Coulee and south of Magrath (Pl. 3). The moraine pattern indicates that there was a distinct Glenwoodville ice lobe, which advanced up the preglacial valley of Oldman River to a point about 5 miles west of Pincher Creek.

The Glenwoodville till, although rarely exposed because of the cover of lake deposits, is essentially similar to the other continental tills.

### *Regional Interpretation and Correlation of Continental Moraines*

The areal relations of the continental moraines within the region show that they were deposited by a major ice lobe that protruded westward from the position of the Lethbridge moraine into the broad, regional lowland between the Porcupine Hills on the north and the Rocky Mountain foothills and Milk River Hills to the south (Pl. 3). Although there are several prominent uplands within the general lowland area, they have not greatly modified the regional moraine pattern. The axis of ice movement was possibly initially diverted westward somewhat by impingement against the Milk River Hills, but the major factor in determining its position was the broad lowland along the headwaters of the St. Mary, Belly, Waterton, and Oldman rivers. At the edge of the foothills the main direction of flow, as indicated by the Glenwoodville lobe, was along the preglacial valley of Oldman River between Pincher Creek and Glenwoodville.

On the basis of the areal Pleistocene geology it may be concluded that: there are two major moraines, the Kimball and Glenwoodville, between the Outer Continental moraine and the Lethbridge moraine; the Kimball is the equivalent of the Okotoks moraine (Johnston and Wickendon, 1931, p. 30; Pl. 8) farther north; and Glenwoodville is equivalent to the prominent moraine east of the Okotoks which passes through Clear Lake (Macleod, Sectional Map No. 65) and extends southward to a point about 20 miles northeast of Macleod.

## Terraces

Stream terraces occur along all of the major valleys of the region, but they were not investigated in detail. Representative elevations, based on altimeter readings and the stream profiles shown by topographic maps, are given in Table 2.

The terraces occur below the level of adjoining drift plains, lacustrine plains, and meltwater channels, and thus postdate the major glacial events of the region. Two or three main terraces with local minor terraces are present in most valleys. They are composed largely of

gravel and represent aggradational fills rather than erosional surfaces.

Along Belly River low terraces, 15–35 feet above stream level, are accordant with the present stream profile, but an upper terrace at 40–70 feet may diverge downstream. A persistent set of low terraces, 15–25 feet above stream level, also is present along the Waterton Valley in conjunction with at least two higher terraces. Along Drywood, Yarrow, and Dungarvan valleys the most prominent terrace remnants are 40–60 feet above drainage.

The higher terraces may be related to aggradation controlled by glacial-lake levels in lower reaches of the valley. The limited data available suggest, in fact, that the 40- to 70-foot terrace along Belly River and the 40- to 60-foot terrace along Drywood Valley are graded to the level of glacial Lake Cardston at about 3900 feet. This interpretation, however, requires the assumption that lower lake levels permitting valley deepening preceded the higher level. Since glacial lake levels depended upon the position of the ice front with respect to outlet channels at different elevations, any major fluctuation would require an important retreat and readvance of the ice. There is no evidence at present of such an event, but it is not improbable. In this connection it may be noted that a major retreat and readvance to the position of the Lethbridge moraine caused the rebirth of Glacial Lake Macleod farther down valley (Horberg, 1952b, p. 316). A record of this event in terms of valley erosion and subsequent aggradation may be provided by terraces in the northeastern part of the area, and certainly the

TABLE 2.—TERRACE ELEVATIONS

| Location | Elevation | Height above drainage |
|---|---|---|
| Belly River: | | |
| (1) SE. ¼, sec. 8, T. 2, R. 28; W. of Payne Lake | 4392 | 42 |
| (2) SE. ¼, sec. 20, T. 2, R. 28; Cardston highway | 4276 | 16 |
| | 4315 | 55 |
| (3) NW. ¼, sec. 35, T. 2, R. 28; Caldwell | 4090 | 35 |
| | 4127 | 72 |
| Waterton River: | | |
| (1) NE. ¼, sec. 24, T. 3, R. 29; mouth of Dungarvan Creek | 4040 | 20 |
| (2) SE. ¼, sec. 30, T. 3, R. 28; 1 mile E. of Dungarvan Creek | 4012 | 7 |
| | 4031 | 26 |
| (3) NW. ¼, sec. 23, T. 4, R. 28; Hill Spring road | 3772 | 12 |
| | 3782 | 22 |
| | 3800 | 40 |
| | 3883 | 123 |
| Drywood River: | | |
| (1) SE. ¼, sec. 21, T. 4, R. 29; 2 miles above mouth of Yarrow Creek | 4292 | 42 |
| | 4352 | 102 |
| (2) NW. ¼, sec. 7, T. 4, R. 28; 2 miles below mouth of Yarrow Creek | 4020 | 20 |
| | 4038 | 38 |
| | 4054 | 54 |
| Yarrow Creek: | | |
| (1) SE. ¼, sec. 8, T. 4, R. 29; 2 miles above mouth of Dungarvan Creek | 4310 | 60 |
| Dungarvan Creek: | | |
| (1) NW. ¼, sec. 8, T. 3, R. 29; 5.5 miles above mouth | 4600 | 20 |
| | 4640 | 60 |
| Blakiston Creek: | | |
| (1) Center, sec. 1, T. 2, R. 30; mouth of Blakiston Canyon | 4375 | 35 |
| | 4410 | 70 |
| | 4435 | 95 |

possibility should be considered in later investigations.

The lower terraces, 10–25 feet above stream levels, in upper Waterton and Dungarvan valleys and probably in other valleys are correlated with alluvial fans and valley gravels which are believed to have been deposited during the postglacial Little Ice Age.

## Glacial Lakes

### *General*

The existence of former water planes in the region was first recognized by Dawson and McConnell (1895, p. 60–63), but successive glacial lakes at progressively lowered levels were not distinguished until much later. Lakes Magrath and Lethbridge were named and delineated by Johnston and Wickendon (1931, p. 41, Fig. 1) and Lakes Cardston and Macleod, and the stages of Lake Magrath, were distinguished by Bretz (1943, p. 38–39; Pl. 3). South of the International Boundary Alden (1932, p. 103–104; Pl. 37) recognized Lake Cut Bank and glacial lakes in South Fork Milk and St. Mary valleys.

In the present study Lake Caldwell, named from the town of Caldwell in the Mountain View quadrangle, is distinguished as a major stage preceding Lake Cardston. Smaller lakes confined to a single valley are recognized in Dungarvan, Waterton, Belly, St. Mary (Lakes St. Mary and Farehard), Milk (North Fork lakes and Lake Twin River), and Red valleys (Pl. 6).

### *Lacustrine Features*

Lacustrine features within the area include: (1) the extensive lake plains of glacial lakes Caldwell, Cardston, and Magrath; (2) various shore features which border the lake plains and in places are present in morainic areas; (3) lake deposits which occur throughout the area at elevations below about 5000 feet; and (4) a complex system of outlet channels at various levels.

The shore features, although not conspicuous, are readily recognizable at many places (Pl. 1). They include wave-cut scarps, beach ridges, beach-ridge lagoons, and wave-cut platforms; the scarps generally are the most prominent. In addition there are distinctive forms resulting from the modification of morainic knobs and basins within the shore zone. The general case, which was observed at several places, shows unmodified moraine above the shore line, a scarp and beach ridge at the shore line, wave-truncated knobs with residual boulders and partially filled basins near shore, and mantled moraine offshore.

Inasmuch as proglacial lakes were present continuously in the major valleys as the continental ice withdrew, lake deposits occur at elevations ranging from over 5000 feet in the early Lake Waterton basin to 3250 feet in the Lake Magrath basin in the northeastern corner of the area. The highest observed occurrence is along the Chief Mountain Highway east of Sofa Creek (NE. ¼, sec. 27, T. 1, R. 29) where pink and buff, varved clays and silts occur at an elevation of 5050 feet (altimeter). The finely laminated or varved lake-bottom deposits are maroon or pink near the mountains, but in the foothills and lowlands they are uniformly gray or buff. In many places, because of the prevalence of steeply sloping surfaces of deposition, they are distorted by penecontemporaneous and later-slump movements. The shore deposits consist mainly of inclined beds of gravel with sand and rare silt layers. Representative exposures of lake deposits can be examined readily at the following localities:

WATERTON VALLEY:

(1) North end of bedrock ridge 1.4 miles northeast of Waterton—varved clays
(2) Crooked Creek, 0.7 mile above mouth (sec. 10, Appendix)—silts
(3) Pincher Creek road, NE. corner, sec. 32, T. 2, R. 29—silts
(4) Cottonwood Creek and Pincher Creek road, NW. ¼, sec. 20, T. 2, R. 29—clays

BELLY RIVER VALLEY:

(1) Cardston highway bridge (sec. 5, Appendix)—thick silts
(2) South of above 1.5 miles (sec. 2, Appendix) —clays

LAKE CALDWELL BASIN:

(1) Hill Spring highway, 1 mile north of Caldwell (SE. ¼, sec. 1, T. 3, R. 28)—silts

(2) Cardston highway near Leavitt (secs. 4, 5, T. 3, R. 26)—silts

LAKE CARDSTON BASIN:

(1) Waterton highway, 1.5 miles west of Cardston (NE. ¼, sec. 7, T. 3, R. 25)—beach gravels
(2) Hill Spring road, 6 miles northeast of Cardston (NE. ¼, sec. 28, T. 3, R. 26)—beach gravels
(3) Cardston–Magrath highway and St. Mary River (sec. 23, T. 3, R. 25)—silts

In places isolated large boulders of Belt quartzite and dolomite, up to 25 feet in diameter, occur on the Cardston lake plain. As suggested by Rutherford (1941), these are probably best explained by ice rafting from till exposures along the shore line.

The outlet valleys of the glacial lakes, and related ice-margin channels are shown in Plates 1 and 6. They vary in character from small, ill-defined channels to large, abandoned valleys and are eroded in both bedrock and glacial drift. Most channels are poorly drained, and many of the larger ones contain lakes.

### *Shoreline Elevations*

Precise data on shoreline elevations are limited. A summary of elevations determined by altimeter is given in Table 3. These elevations and additional elevations determined from topographic maps indicate that the main stages of Lakes Caldwell and Cardston were at elevations of about 4225 and 3900 feet respectively, and that there is no evidence of post-glacial tilting of the shore lines.

Shoreline features are rarely observable above the Lake Caldwell level, but in places there are traces of shore lines of the earlier lakes and of possible intermediate stages. Noteworthy examples are the traces of Lake Waterton shore lines at three or four levels on the bedrock ridge on the east side of the valley and scarps of a possible pre-Caldwell lake at elevations of about 4630 feet in the foothills 8 miles south of Pincher Creek.

Possible higher stages of Lake Cardston are evidenced by scarps and beach gravels at elevations of 4000–4100 feet south and west of Cardston.

### *Glacial Lake History*

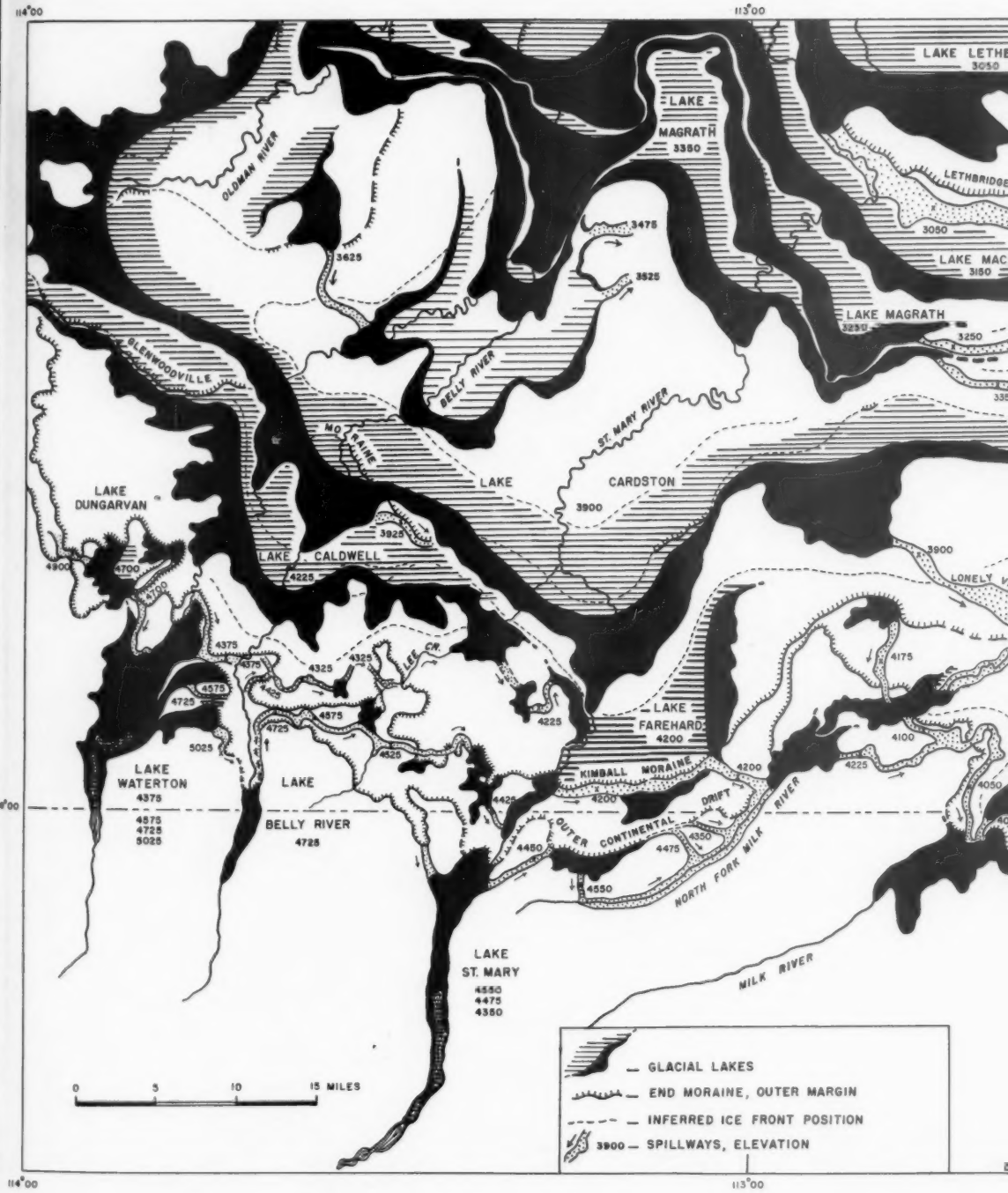
The complex history of ice-margin lakes and drainage lines is summarized in the light of existing information for the region as a whole in Table 4 and Plate 6. There are many uncertainties in correlations, and numerous details of intermediate stages remain to be filled in, but the major events are clearly recorded from the evidence at hand. The evidence for the later stages consists of shoreline features, outlet channels, and the lake plains themselves, but that for the earlier stages is restricted largely to ice-margin and outlet channels, scattered occurrences of lake silts, and the moraines from which the position of ice dams can be inferred. In most places the shore lines shown in Plate 6 were drawn by extending contours of outlet heads back into the lake basins.

No attempt will be made to describe the lake stages in detail, but some of the important interpretations will be considered briefly, beginning with the earlier stages characterized by valley lakes.

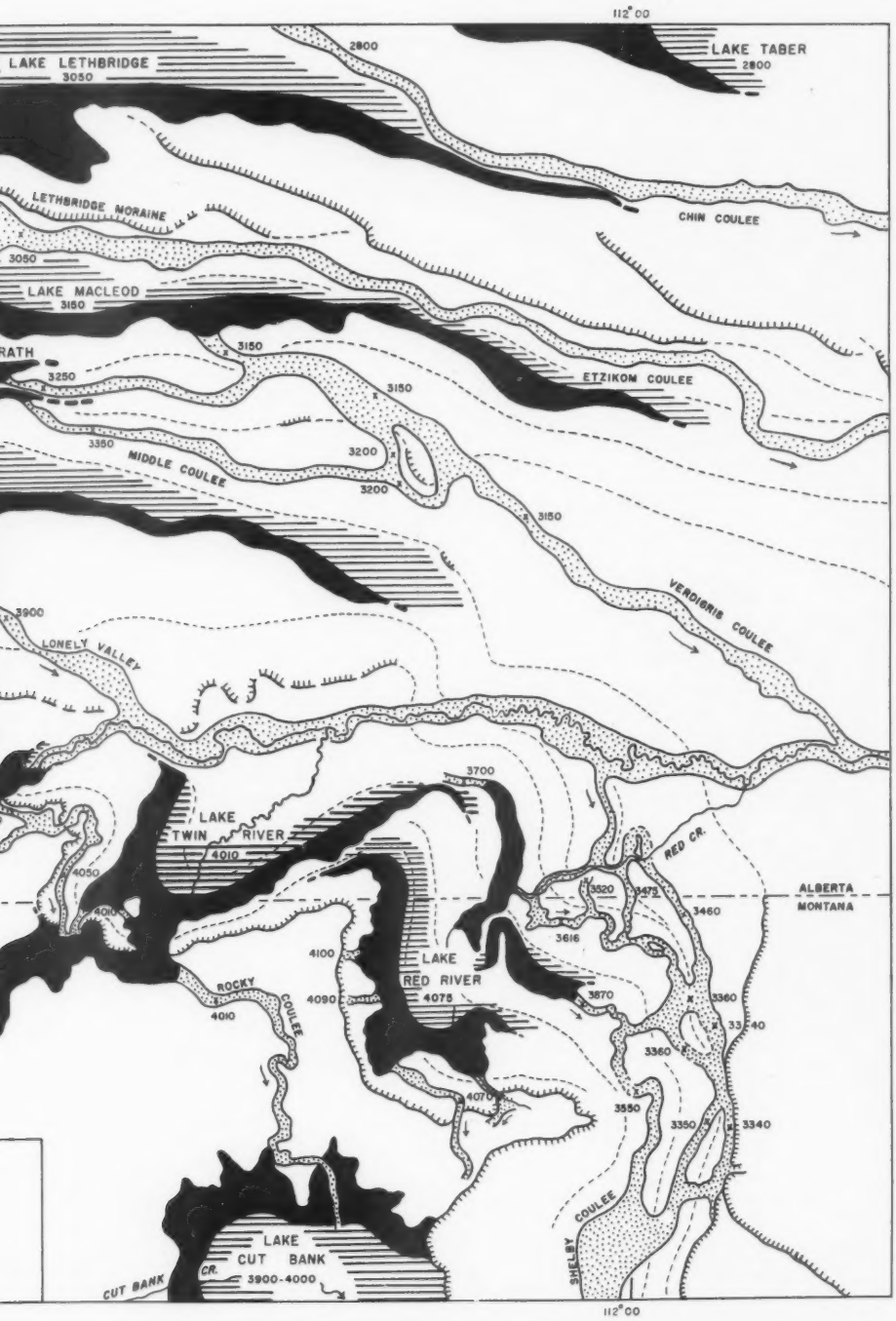
The lake stages in the Waterton and Belly River valleys are inferred mainly from ice-front positions and channels which record continuous lakes from the time of maximum advance of the ice until Kimball retreat. During the earliest stage of maximum advance the lakes must have discharged eastward along the edge of the ice and probably through subglacial channels, but there is no evidence of a continuous ice-margin channel. At a later stage the prominent channel at 4575 feet (Pl. 6) was occupied, and during final stages the drainage appears to have been in part subglacial by way of the channel at 4325 feet into Lee Creek.

Early lake stages in the St. Mary Valley were determined by successively lower outlets at elevations of about 4550, 4475, and 4350 feet across Hudson Divide into North Fork Milk River. During later stages the Mary Lake channel at 4200 feet was utilized. From St. Mary Valley the drainage was eastward into North Fork Lake, thence into early Lake Twin River, and finally by way of Rocky Coulee into Lake Cut Bank.

With the opening of the Lake Caldwell stage, the valley lakes largely disappeared, and the proglacial waters expanded into lowland lakes which were no longer confined to a single drain-




GLACIAL GEOLOGY OF THE WATERTON




VATERTON REGION

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age basin. Lake Caldwell drained eastward into Lake Farehard, and from the Lake Farehard outlet there was probably essentially free drainage by way of Milk River into successive channels leading into Shelby Coulee.

the Glenwoodville with a younger moraine which overlaps the older a short distance to the north (Pl. 6).

In conclusion it should be emphasized that the record is one of continuous proglacial lakes

TABLE 3.—ELEVATIONS OF SHORE LINES OF GLACIAL LAKES

| Location | Shore feature | Elevations |
|---|---|---|
| Lake Caldwell: | | |
| Caldwell, 0.75 miles SE. (NE. ¼, sec. 25, T. 2, R. 28) | scarp | 4227 |
| Caldwell, 5 miles NW. (NW. ¼, sec. 16, T. 3, R. 28) | scarp | 4225* |
| Leavitt, 1.25 miles W. (SE. ¼, sec. 6, T. 3, R. 26) | scarp | 4226 |
| Hill Spring, 7 miles W. (SE. ¼, sec. 24, T. 4, R. 28) | scarp | 4238 |
| Hill Spring, 8 miles W. (NW. ¼, sec. 24, T. 4, R. 28) | channel | 4235 |
| Lake Cardston: | | |
| Cardston, 1.5 miles W. (NW. ¼, sec. 7, T. 3, R. 25) | beach ridge | 3910 |
| Cardston, 5 miles NW. (sec. 22, T. 3, R. 26) | scarp | 3900 |
| Hill Spring, 3.5 miles NW. (NE. ¼, sec. 22, T. 4, R. 28) | scarp | 3922 |
| Hill Spring, 5 miles NW. (NW.¼, sec. 34, T. 4, R. 28) | beach ridge | 3957 |
| Pincher Creek, 1.5 miles S.SE. (NE. ¼, sec. 11, T. 6, R. 30) | scarp | 3861 |
| Pincher Creek, 1 mile S. (SW. ¼, sec. 14, T. 6, R. 30) | scarp | 3851 |
| Pincher Creek, 1 mile SW. (NW. ¼, sec. 15, T. 6, R. 30) | scarp | 3855 |
| Pincher Creek, 1.5 miles W. (SE. ¼, sec. 2, T. 6, R. 30) | scarp | 3913 |
| Pincher Station, 0.5 miles N. (NE. ¼, sec. 2, T. 7, R. 30) | scarp | 3898 |
| Cowley, 2 m. S. (SE. ¼, sec. 9, T. 7, R. 1) | scarp | 3900* |

* Topographic map reading

As the ice withdrew still farther, the waters of Lake Cardston expanded widely throughout the regional lowland into all the major drainage basins of the region. This local maximum flooding was followed in the Lake Magrath stages by lakes of more restricted extent in the Oldman and Waterton-Belly drainage basins. Intermediate stages between Lake Cardston and the Lake Magrath stages of Bretz (1943, Pl. 3) are represented by stages at an elevation of 3625 feet in the Oldman drainage basin and at elevations of 3525 and 3475 feet in the Waterton-Belly drainage basin. The main stages of Lake Magrath and the stages of Lakes Macleod, Lethbridge, and Taber, which followed, are clearly recorded by lake plains and impressive outlet coulees.

Correlation of lake stages and their relations to ice-front positions indicated by moraines are given in Table 4 and shown in Plate 6. The regional relations thus suggest tentative correlations of the Kimball moraine with the older moraine on the east side of Shelby Coulee and at progressively lower levels, and the lake stages were of such relatively brief duration that prominent shore features were developed in only a few places.

## DRIFT LITHOLOGIES

Pebbles half an inch to 1½ inches in diameter were used as an index grade size for determining lithologies of the various drifts, and collections of 100 pebbles were taken from the surface of exposures. The major results, based on data given in Table 5 and shown graphically in Figure 5, may be summarized as follows:

(1) The mountain till, compared with the continental till, is characterized by a higher content of argillite (average, 56.5% *vs.* 20.5%) and lower percentages of quartzite (13.5% *vs.* 20.3%), limestone and dolomite (20.3% *vs.* 36.2%), and Cretaceous shale and sandstone (3.8% *vs.* 12.1%). Precambrian crystallines are absent from the mountain till but average 3 per cent in the continental till.

TABLE 4.—HISTORY OF GLACIAL LAKES

| Moraine sequence | Oldman River | Waterton River | Belly River | St. Mary River | North Fork Milk River | Milk River | Red River | Cut Bank Creek |
|---|---|---|---|---|---|---|---|---|
| | Lakes—Elevations | Lakes—Elevations | Lakes—Elevations | Lakes—Elevations | Lakes—Elevations | Lakes—Elevations | Lakes—Elevations | Lakes—Elevations |
| Buffalo Lake | Lake Taber—2800 (Chin Coulee outlet) | | | | | | | |
| Chin | Lake Lethbridge—3050 (Etzikom Coulee) | | | | | | | |
| Lethbridge | Lake Macleod—3150 (Verdigris Coulee) | | | | | | | |
| | Lake Magrath—3250 (Verdigris Coulee) | Lake Magrath—3250 | Lake Magrath—3250 | | | | | |
| Glenwoodville retreat | Lake Magrath—3350 (Middle Coulee) | Lake Magrath—3350 | Lake Magrath—3350 | | | | | |
| | Lake Magrath—3625 | Lake Magrath—3475<br>Lake Magrath—3525 | Lake Magrath—3475<br>Lake Magrath—3525 | | | | | |
| Glenwoodville maximum | Lake Cardston—3900 (Lonely Valley) | Lake Cardston—3900 | Lake Cardston—3900 | Lake Cardston—3900 | | | | |
| | Lake Caldwell—4225 (N. Fork Milk | Lake Caldwell—4225 | Lake Caldwell—4225 | Lake Farehard—4200 (Milk River- | | | | |

| | | | | | | | | |
|---|---|---|---|---|---|---|---|---|
| Kimball retreat | Lake Caldwell —4225 (N. Fork Milk River) | Lake Caldwell —4225<br>Lake Waterton —4375 | Lake Caldwell —4225<br>local lake—4325<br>Lake Belly River—4375<br>Lake Belly River—4425 | Lake Farehard —4200 (Milk River-Shelby Coulee) | | | | |
| Kimball maximum | | Lake Waterton —4575<br>Lake Dungarvan—4700 | Lake Belly River—4575 | Lake St. Mary—4200 (Mary Lake channel, 4200) | local lake—4175 | Lake Twin River—3700 (Rocky Coulee) | Lake Red River —3616 (Shelby Coulee) | |
| Outer Continental retreat | | Lake Waterton —4575<br>Lake Waterton —4725 (Birdseye Butte) | Lake Belly River—4575<br>Lake Belly River—4725 | Lake St. Mary—4350<br>Boundary Creek Lake—4425<br>Lake St. Mary—4475 | valley lake—4100 | | Lake Red River —3870<br>Lake Red River —4075 (Shelby Coulee) | |
| Outer Continental maximum | | (Birdseye Butte)—5025 (col to Belly River) | | Lake St. Mary—4550 (N. Fork Milk River) | valley lake—4225 (Rocky Coulee) | Lake Twin River —4010 (Rocky Coulee) | | Lake Cut Bank —3900–4000 (Cut Bank Creek) |

TABLE 5.—DRIFT LITHOLOGIES (Number per cent of index grade size)

| Localities (*see* Fig. 5) | Belt series | | | | | | | | | Largely belt | | ? | | Cretaceous | | Precambrian shield | | | Belt *summary* | | Largely belt *summary* |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Quartzite | | | | Argillite | | | Basalt | Andesite porphyry | Dolomite | Limestone | Chert | Quartz | Sandstone | Shale | Granitic | Gneiss, schist | Basic | Quartzite | Argillite | Limestone-dolomite |
| | Purple-red | White-buff | Dense gray | Coarse gray | Purple-red | Green-gray | Buff | | | | | | | | | | | | | | |
| Late Mountain drift: | | | | | | | | | | | | | | | | | | | | | |
| Waterton moraine | | | | | | | | | | | | | | | | | | | | | |
| 1. Kame gravel—Waterton Park | 6 | 19 | 1 | .. | 15 | 21 | 16 | 2 | 11 | 2 | 7 | .. | | .. | .. | | | | 26 | 52 | 9 |
| 2. Till—East of Lower Waterton Lake | .. | 2 | 1 | 1 | 20 | 44 | 5 | .. | .. | 11 | 2 | 3 | | 11 | .. | | | | 4 | 69 | 13 |
| 3. Till—Galwey Creek | 2 | 16 | .. | 1 | 12 | 37 | 6 | 2 | 2 | 10 | 12 | .. | | .. | .. | | | | 19 | 55 | 20 |
| 4. Till—Crooked Creek area | 3 | 8 | 2 | .. | 15 | 27 | 6 | .. | .. | 24 | 1 | 6 | | 5 | 2 | | | | 13 | 48 | 24 |
| 5. Till—Crooked Creek area | 5 | 7 | .. | 5 | 7 | 24 | 11 | .. | 3 | .. | 29 | 6 | | .. | 3 | | | | 17 | 42 | 30 |
| 6. Till—Crooked Creek area | 1 | 15 | .. | .. | 12 | 28 | 14 | .. | 2 | 16 | 1 | 7 | | 4 | .. | | | | 16 | 54 | 17 |
| 7. Till—Crooked Creek area | 2 | 6 | 1 | 3 | 8 | 28 | .. | .. | .. | 12 | 36 | .. | | 3 | .. | | | | 12 | 36 | 48 |
| Till average, Waterton moraine | 2.2 | 9.0 | 0.7 | 1.3 | 12.3 | 31.7 | 7.0 | 0.3 | 1.2 | 12.2 | 13.5 | 3.7 | | 3.0 | 0.8 | | | | 13.5 | 50.7 | 25.2 |
| Sofa Creek moraine | | | | | | | | | | | | | | | | | | | | | |
| 8. Till—Chief Mtn. Hwy. | .. | 5 | .. | .. | 12 | 45 | 21 | .. | .. | .. | 14 | .. | | 3 | .. | | | | 5 | 78 | 14 |
| Belly River moraine | | | | | | | | | | | | | | | | | | | | | |
| 9. Till—Crooked Creek basin | .. | 21 | .. | .. | 7 | 46 | 13 | 1 | 1 | 8 | .. | 2 | | .. | .. | | | | 21 | 66 | 8 |
| 9. Till—Crooked Creek basin | .. | 22 | .. | .. | 9 | 57 | 6 | .. | .. | .. | 2 | .. | | 2 | 1 | | | | 22 | 71 | 2 |
| 10. Till—Lookout Butte | .. | 15 | .. | .. | 11 | 34 | 32 | .. | 1 | 1 | 1 | 2 | | 2 | 1 | | | | 15 | 77 | 2 |
| Average, Belly River till | .. | 19.3 | .. | .. | 9.0 | 34.2 | 17.0 | 0.3 | 0.7 | .. | .. | 1.3 | | 1.3 | 0.7 | | | | 19.3 | 71.3 | 4.0 |
| Mountain drift below Continental drift: | | | | | | | | | | | | | | | | | | | | | |
| 11. Till—Waterton River | 1 | 14 | 1 | 1 | 10 | 31 | .. | 3 | 3 | 32 | 2 | 1 | | .. | .. | | | | 17 | 41 | 34 |

| | | | | | | | | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Average, Belly River till | .. | 19.3 | .. | .. | 9.0 | 34.2 | 17.0 | 0.3 | 0.7 | .. | .. | 1.3 | | 1.3 | 0.7 | | | | 19.3 | 71.3 | 4.0 |

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|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Mountain drift below Continental drift: | | | | | | | | | | | | | | | | | | | | | | |
| 11. Till—Waterton River | 1 | 14 | 1 | 1 | 10 | 31 | .. | 3 | 3 | 32 | 2 | 1 | | .. | .. | | | | 17 | 41 | 34 |
| 12. Gravel—Belly river bridge | 10 | 15 | ..43.. | | 2 | 4 | 12 | .. | .. | .. | 14 | .. | | .. | .. | | | | 68 | 18 | 14 |
| 13. Till — Belly River — Caldwell | 6 | 10 | 8 | .. | 8 | 26 | .. | 1 | 2 | 26 | 6 | .. | | 7 | .. | | | | 24 | 34 | 32 |
| Till average, 11 and 13 | 3.5 | 12.0 | 4.5 | 0.5 | 9.0 | 28.5 | .. | 2.0 | 2.5 | 29.0 | 4.0 | 0.5 | | 3.5 | .. | | | | 20.5 | 37.5 | 33.0 |
| Early Mountain drift: | | | | | | | | | | | | | | | | | | | | | | |
| 14. Till—Sofa Creek | .. | 6 | .. | 3 | 15 | 46 | 3 | .. | 1 | 21 | .. | .. | | 6 | .. | | | | 9 | 64 | 21 |
| Average—All mountain till | .. | .. | .. | .. | 11.2 | 36.3 | 9.0 | 0.5 | 1.2 | 14.7 | 8.2 | 2.1 | | 3.3 | 0.5 | | | | 14.9 | 56.5 | 20.3 |
| Continental drift: | | | | | | | | | | | | | | | | | | | | | | |
| 12. Till—Belly River bridge | 4 | 26 | .. | 3 | 1 | 12 | .. | .. | .. | 11 | 16 | 9 | .. | 18 | .. | .. | .. | .. | 33 | 13 | 27 |
| 15. Upper till—Belly River bridge | 1 | 12 | .. | .. | 10 | 22 | 4 | 1 | 1 | 20 | 2 | 3 | .. | 20 | 3 | 1 | .. | .. | 13 | 36 | 22 |
| 15. Lower till—Belly River bridge | 5 | 8 | .. | 2 | .. | 16 | 11 | .. | .. | 45 | 2 | 1 | .. | 6 | 4 | .. | .. | .. | 15 | 27 | 47 |
| 16. Till—Cardston Road | 1 | 5 | .. | .. | 14 | 29 | 6 | .. | 1 | 3 | 18 | 7 | .. | 11 | .. | 1 | 1 | 2 | 6 | 49 | 21 |
| 13. Gravel—Belly River—Caldwell | 6 | 12 | 3 | .. | 2 | 16 | 1 | .. | .. | 13 | 12 | 8 | .. | 14 | 6 | 6 | 1 | .. | 21 | 19 | 25 |
| 17. Gravel—Lookout Butte | 3 | 10 | 1 | .. | 2 | 14 | 17 | .. | .. | 12 | 5 | 4 | 2 | 16 | 11 | 3 | .. | .. | 14 | 33 | 17 |
| 11. Till—Waterton River | 6 | 18 | .. | 5 | 11 | 10 | 2 | 1 | 1 | 9 | 23 | 4 | .. | 5 | .. | 4 | 1 | .. | 29 | 23 | 32 |
| 19. Till—Mouth of Dungarvan Creek | 3 | 15 | .. | .. | 2 | 1 | 2 | .. | .. | ..49.. | | 9 | .. | 13 | .. | 6 | .. | .. | 18 | 5 | 49 |
| 18. Till—Dungarvan Creek | .. | 15 | 1 | .. | 1 | 2 | 3 | 1 | .. | 30 | 14 | 1 | .. | 26 | .. | 6 | .. | .. | 15 | 6 | 44 |
| 20. Till—Dungarvan Creek | 2 | 13 | 4 | 1 | .. | 1 | .. | .. | 1 | ..55.. | | 6 | 1 | 12 | .. | 4 | .. | .. | 20 | 1 | 55 |
| 21. Till—Dungarvan Creek | 2 | 27 | 3 | 4 | 10 | 18 | 4 | 3 | 4 | ..14.. | | 3 | 1 | 5 | .. | 1 | .. | .. | 36 | 32 | 14 |
| 22. Till—Dungarvan Creek | 1 | 16 | .. | 1 | 3 | 2 | 8 | 2 | .. | 4 | 47 | 5 | | 8 | .. | 3 | .. | .. | 18 | 13 | 51 |
| Average—Continental till | 2.5 | 15.5 | 0.8 | 1.6 | 5.2 | 11.3 | 4.0 | 0.8 | 0.8 | .. | .. | 4.8 | 0.2 | 11.4 | 0.7 | 2.6 | 0.2 | 0.2 | 20.3 | 20.5 | 36.2 |
| Average—Continental gravel | 4.5 | 11.0 | 2.0 | .. | 2.0 | 15.9 | 9.0 | .. | .. | 12.5 | 8.5 | 6.0 | 1.0 | 15.0 | 8.5 | 4.5 | 0.5 | | 17.5 | 26.0 | 21.0 |

(2) The high percentages of Belt rocks compared with local Cretaceous sandstone and shale and foreign constituents in the continental drift show that much of the drift was derived locally from underlying mountain drift. Observations in the field, however, indicate that higher percentages of Cretaceous material are present in finer grade sizes and because of low durability do not show up as pebbles in the index grade.

(3) The fact that argillite in the mountain drift decreases in abundance away from the mountains and is less abundant than other Belt constituents in the continental drift is probably due to its low durability.

(4) Limestone and dolomite ratios are erratic, and their combined percentages vary from place to place in both continental and mountain drifts. At locality 7 (Fig. 5) the high content in mountain drift is due to local limestone ledges, but the reasons for high percentages in continental drifts at localities 18, 19, 20, and 22 are unknown. Most carbonate pebbles are from the Belt series, where limestones and dolomites are abundant and have about equal total stratigraphic thicknesses, but a few may be derived from distant Paleozoic sources.

(5) Quartzite, because of its high durability, is abundant in the continental drift as a rehandled constituent derived from underlying mountain drift. The unusually high content (68%) in the mountain outwash at locality 12 also reflects quartzite durability and its increase in relative abundance with transport.

(6) Variation in the lithology of mountain till from canyon to canyon is illustrated by a comparison of Sofa Creek till (localities 8, 14), in which limestone and dolomite are second in abundance, with Belly River till (locality 10), in which quartzite is second in abundance and limestone and dolomite are rare. On this basis the till at locality 9, whose derivation was somewhat in doubt, is clearly derived from Belly River rather than from Sofa Creek—which confirms the geomorphic evidence.

(7) At localities (11, 12; and 13) where continental drift overlies mountain drift there are notable variations between the two, but they are different at each locality. At locality 11 there is a decrease in argillite and an increase in quartzite in the continental drift; at 12 there is a decrease in quartzite and argillite and an increase in Cretaceous rocks and limestone and dolomite; and at 12 there is a decrease in argillite and an increase in Cretaceous rocks.

The pebble counts can be correlated with the field properties of the mountain and continental tills only in a general way. The abundant purplish-red and greenish-gray argillites are clearly reflected in the purple gray color of the mountain tills. However, the dark gray color of the continental tills is due not so much to the decrease in argillite and increase in other constituents as to the increase in proportions of finer matrix composed in large part of Cretaceous shale.

## Soil Profiles

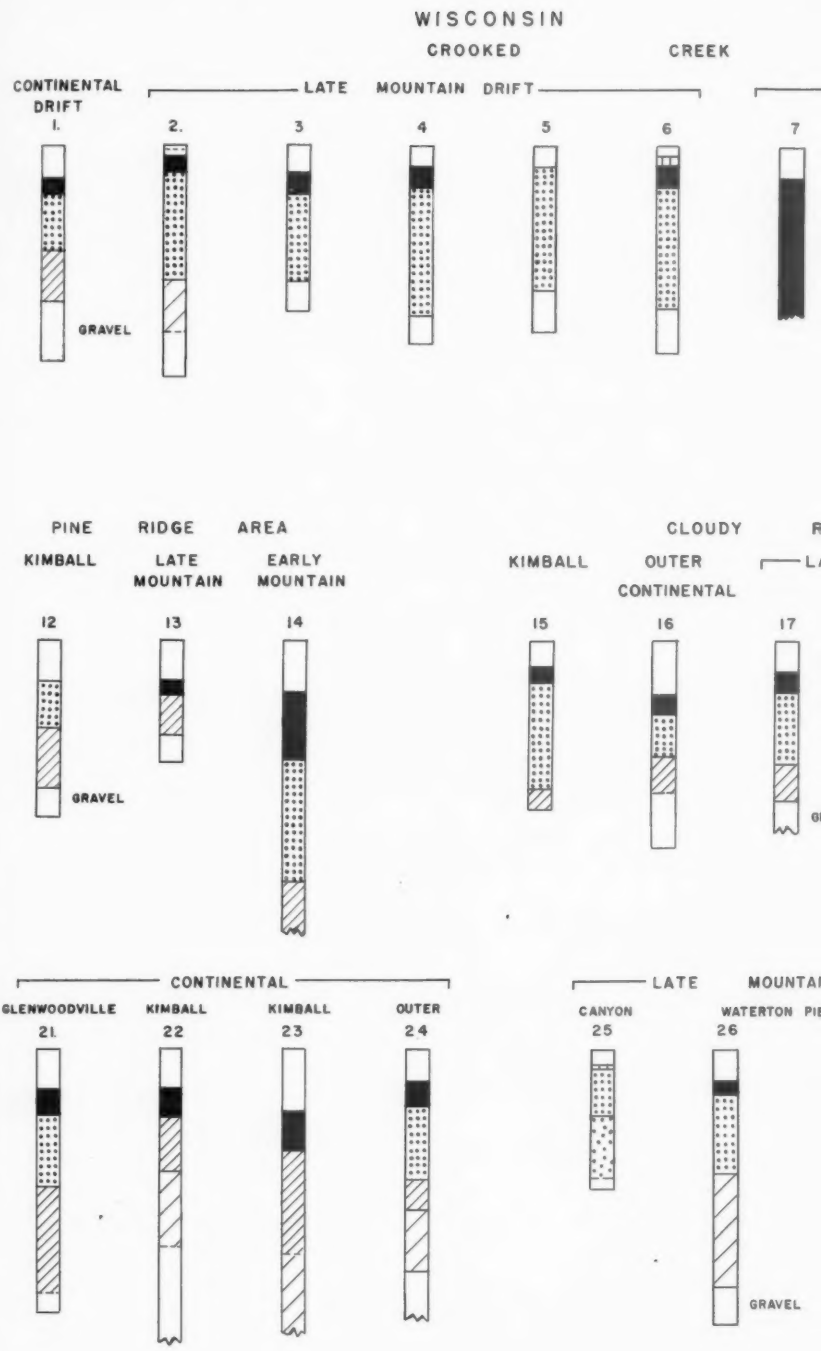
### *Soil Types*

The soils studied lie largely within the foothill zone of transition from pedocal to podzolic types. They include black and shallow black Chernozem and Prairie soils in grassland areas and gray and brown podzolic soils on forested slopes and in the canyons. Almost all profiles measured were developed on glacial tills, including both mountain and continental types.

### *Zonal Terminology*

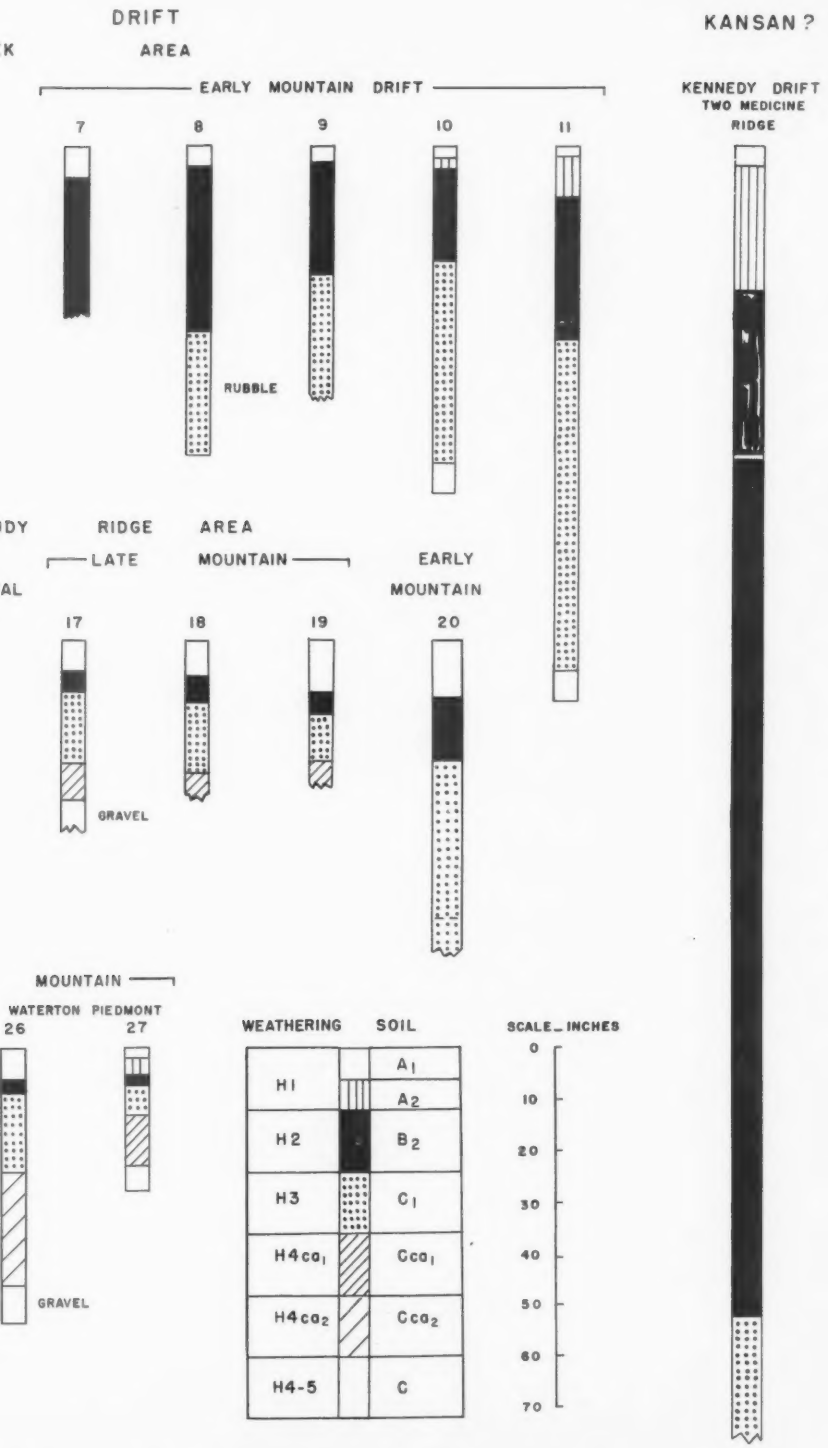
The horizon designations recognized in the present study represent an attempt to reconcile the horizons of the soil profile adopted by soil scientists (Soil Survey Manual, 1951, 174–183) with the horizons of the profile of weathering of Leighton and MacClintock (1930), which are used generally by Pleistocene geologists in the Mississippi Valley (Chart, Pl. 7). This resulted in modifications of both terminologies: the horizon of lime accumulation, *H*4ca, is recognized as a subdivision of the *H*4 calcareous oxidized horizon in the profile of weathering; and the leached and oxidized horizon, *H*3, is designated $C_1$ in the soil profile. The major characteristics of the horizons are as follows:

*H*1 *A* —horizon of organic accumulation and clay depletion; loose, friable
$A_1$ —horizon of surficial organic accumulation in podzolic soils
$A_2$ —lighter-colored eluvial horizon in podzolic soils




SOIL PROFILES ON GLACIAL DEPOSI

For locations and descriptions see Plat




CIAL DEPOSITS IN WATERTON REGION

IONS SEE PLATE 8, TABLE 6, AND APPENDIX A.

*H2* *B*

*H3* $C_1$

*H4* *C*
*H4ca* *Cca*
*H4ca*$_1$*Cca*

*H4ca*$_2$*Cca*

*H5* *C*

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*H*2 *B* —horizon of clay accumulation; textural *B* or $B_2$ horizon of soil scientists; gumbotil, siltil, or mesotil in mature profiles, as designated by geologists; plastic, hard

*H*3 $C_1$ —leached and oxidized horizon; crystalline rocks partially decomposed; friable

*H*4 *C* —calcareous oxidized horizon

*H*4ca *C*ca —lime-enriched horizon

*H*4$ca_1$*C*$ca_1$—lime disseminated throughout matrix and coats all particles; close-set veinlets

*H*4$ca_2$*C*$ca_2$—lime dissemination partial and only some particles are coated; widely spaced veinlets

*H*5 *C* —unaltered parent material

### *Method of Study*

Profiles were measured, insofar as possible, in areas where drifts of different ages are in close proximity and there are no essential differences in climate. The profiles are thus grouped into three areas: the Crooked Creek area, which is 4 miles long and 1 mile wide; the Pine Ridge area, which is 2 miles long and less than 1 mile wide; and the Cloudy Ridge area, which is 3 miles long and less than 1 mile wide (Pls. 7, 8). The measurements were made on the low slopes of well-drained ridge and hill crests and in most cases with grass as vegetation and till as parent material (Table 6). Both exposed sections and auger borings were utilized.

### *Comparison of Profiles*

*General.*—Differential development of profiles on drifts of different ages are clearly indicated, and a chronosequence of three soil types may be recognized: (1) a deep old-age or paleosol profile on the pre-Wisconsin drift; (2) a moderate, mature profile on the Early Wisconsin mountain drift; and (3) shallow, youthful profiles on the Late Wisconsin mountain and continental drifts (Pl. 7). Detailed descriptions of representative profiles are given in Appendix A, and measurements are summarized in Table 6.

*Profiles on Pre-Wisconsin Kennedy drift.*—The paleosol shown by exposed sections of the Kennedy drift on Makowan Butte has been truncated so that only the lower portion of the profile including Horizons 3, 4ca, and 5 can be observed. Horizon 3 consists largely of quartzite, argillite, and a few, partly decomposed diorite pebbles and boulders in a limonitic clayey matrix. Except for solution-etched remnants of dolomite boulders, which were noted at a few places, the zone is thoroughly leached. Below Horizon 3 are prominent ledges of lime-cemented tillite and conglomerate up to 10 feet thick which are believed to represent the horizon of lime enrichment (*H*4ca). The persistent occurrence of the indurated horizon at a uniform level below thoroughly leached deposits supports this view and opposes the alternative interpretation of induration by ground water. The deposits below the indurated horizon are poorly exposed, but in a few places the calcareous, oxidized, nonindurated till in the lower part of Horizon 4 can be observed.

Paleosols on the Kennedy drift are more nearly completely preserved and better exposed on St. Mary Ridge and Two Medicine Ridge in Glacier Park (Browning quadrangle). On St. Mary Ridge (center, sec. 11, T. 34 N., R. 14 W.) the truncated Horizon 3 is 38 feet thick and overlies 15 feet of lime-cemented tillite, the base of which is not exposed. The most nearly complete profile observed, however, is on Two Medicine Ridge at the old Lookout Station just east of Highway 2 (SW. ¼, sec. 16, T. 32 N., R. 13 W.) where the truncated gumbotil of Horizon 2 with superposed Recent soil is 19 feet thick (Pl. 7), Horizon 3 is 42 feet thick, Horizon 4ca is 98 feet thick and the parent materials of Horizons 4 and 5 are over 30 feet thick and include till, outwash, and varved clays. This remarkable section cannot be considered fully in the present report, but the described section is given in Appendix A.

*Profiles on Early Wisconsin mountain drift.*—The soils on Early Wisconsin mountain drift, unlike the Kennedy paleosols, clearly fall within the category of modern soils studied by soil scientists. They are characterized by well-developed textural *B* horizons, which range in thickness from 12 to 32 inches and average 22.7 inches, and by total depths of leaching, which range from 47 inches to 82 inches and average 63.7 inches in the complete profiles studied (Table 6). There is a slight tendency toward somewhat stronger colors (chroma) in the *B* horizons than in the younger soils, but the evidence is not conclusive. In most places the *B* horizon is brown to strong brown according

TABLE 6.—SUMMARY OF SOIL-PROFILE OBSERVATIONS

| Location and locality number (*see* Pl. 8) | Parent material | | Vegetation | Major soil type | *A* Horizon* | | *B* Horizon | | $C_1$ Zone Thickness (Inches) | Cca Zone Thickness (Inches) | Depth of leaching |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | Age | Type | | | Thickness (Inches) | Color† | Thickness (Inches) | Color | | | |
| *Crooked Creek area:* | | | | | | | | | | | |
| 1. Chief Mtn. Hwy.; 1.2 mi. NW. of Belly River bridge | Early Continental | Gravel | Grass; aspen | Chernozem | 6 | 10YR2/2 | 3 | 10YR3/4 | 11 | 10 | 20 |
| 2. Chief Mtn. Hwy.; 2.2 mi. W.NW. of bridge | Late Wis. Mtn. | Till | Grass | Chernozem | 2 | 10YR3/2 | 3 | 10YR4/4 | 21 | 10 | 26 |
| 3. Chief Mtn. Hwy.; at Crooked Creek | Late Wis. Mtn. | Till | Grass | Prairie | 5 | 10YR3/2 | 4 | 10YR4/4 | 17 | ..... | 26 |
| 4. Crooked Creek east lateral moraine; 0.6 mi. SW. of 3 | Late Wis. Mtn. (Crooked Cr.) | Till | Grass | Prairie | 4 | 10YR2/1 | 4 | 10YR4/4 | 25 | ..... | 35 |
| 5. Sofa Creek area; 2 mi. SW. of 3 | Late Wis. Mtn. (Sofa Cr.) | Till | Grass | Prairie | 4 | 10YR2/2 | .. | ..... | 24 | ..... | 28 |
| | | | | | | | | | 10YR4/3 | | |
| 6. Chief Mtn. Hwy.; 2 mi. W. of Crooked Cr. | Late Wis. Mtn. (Sofa Cr.) | Till | Pine | Podzolic | $A_1$-2 | 10YR2/1 | 4 | 10YR4/4 | 24 | ..... | 32 |
| | | | | | $A_2$-2 | | | | | | |
| 7. Lookout Butte; 0.7 mi. E. of 3 | Early Wis. Mtn. | Till | Grass | Chernozem | 6 | 10YR2/1 | 20+ | 10YR4/4 | | | 26+ |
| 8. Rubb le ridge; 0.2 mi. SE. of 4 | Early Wis. Mtn. | Rubble | Grass; aspen | Prairie | 4 | 2.5YR2/0 | 32 | 7.5YR4/4 | 24+ | | 60+ |
| 9. Sofa Creek area; 0.1 mi. N. of 5 | Early Wis. Mtn. | Till | Grass; pine | Prairie | 3 | 10YR2/2 | 22 | 10YR5/6 | 23+ | | 48+ |
| 10. Same as 6 | Early Wis. Mtn. | Till | Pine | Podzolic | $A_1$-2 | 10YR2/1 | 18 | 10YR4/4 | 20 | ..... | 62 |
| | | | | | $A_2$-2 | | | | | | |
| 11. Sofa Cr., W. Early Wis. Mtn. lateral; 1.5 mi. W. of 6 and 10 | Early Wis. Mtn. | Till | Pine | Podzolic | $A_1$-2 | 10YR2 2 | 28 | 7.5YR5/6 | 44 | ..... | 82 |
| | | | | | $A_2$-8 | 10YR4/2 | | | | | |
| *Pine Ridge area:* | | | | | | | | | | | |
| 12. Hwy. W. end of Pine Ridge (NW.¼ sec. 3, T. 3, R. 29) | Kimball | Gravel | Grass | Chernozem | 8 | 10/YR2/2 | .. | 10YR3/4 | 9 | 12 | 17 |
| 13. Hwy. 1 mi. S. of 12 (NW. ¼ sec. 33, T. 2, R. 29) | Late Wis. Mtn. (Waterton) | Till | Grass | Chernozem | 7 | 10YR2/2 | 3 | 7.5YR5/4 | ..... | 8 | 10 |
| 14. Pine Ridge, SW. end (SW | Early Wis. Mtn. | Till | Grass | Chernozem | 10 | 7.5YR2/1 | 13 | 7.5YR4/4 | 24 | 10+ | 47 |

| | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| 13. Hwy. 1 mi. S. of 12 (NW. ¼ sec. 33, T. 2, R. 29) | Late Wis. Mtn. (Waterton) | Till | Grass | Chernozem | 7 | 10YR2/2 | 3 | 7.5YR5/4 | ..... | 8 | 10 |
| 14. Pine Ridge, SW. end (SW ¼ sec. 15, T. 3, R. 29) | Early Wis. Mtn. | Till | Grass | Chernozem | 10 | 7.5YR2/1 | 13 | 7.5YR4/4 | 24 | 10+ | 47 |
| *Cloudy Ridge area:* | | | | | | | | | | | |
| 15. Road, 0.8 mi. E of Yarrow Cr. (NW. ¼ sec. 23, T. 3, R. 30) | Kimball | Till | Grass | Chernozem | 5 | 10YR3/2 | 3 | 10YR4/4 | 21 | 3+ | 29 |
| 16. N. end, Cloudy Ridge (SE ¼ sec. 22, T. 3, R. 30) | Early Continental | Till | Grass | Chernozem | 10 | 5YR2/2 | 4 | 7.5YR4/4 | 8 | 7 | 22 |
| 17. Yarrow Valley W. of Cloudy Ridge (SW ¼ sec. 15, T. 3, R. 30) | ? Late Wis. Mtn. | Gravel | Grass | Chernozem | 6 | 5YR2/2 | 4 | 5YR3/2 | 14 | 7+ | 24 |
| 18. Yarrow Valley, 1 mi. S-SW. of 17 | Late Mountain | Till | Grass | Chernozem | 7 | 7.5YR3/2 | 5 | 7.5YR4/4 | 14 | 4+ | 26 |
| 19. Cloudy Ridge (SE. ¼ sec. 15, T. 3, R. 30) | Late Mountain | Till | Grass | Chernozem | 10 | 5YR2/2 | 4 | 7.5YR4/4 | 9 | 5+ | 23 |
| 20. Same as 19. | Early Wis. Mtn. | Till | Grass | Chernozem | 11 | 5YR2/2 | 12 | 7.5YR5/6 | 39+ | | 62+ |
| *Additional profiles:* | | | | | | | | | | | |
| 21. Hwy. 6, 3.7 mi. S. of Pincher Creek (NW. ¼ sec. 36, T. 5, (R. 29) | Glenwoodville | Till | Grass | Chernozem | 8 | 10YR2/1 | 5 | 10YR4/4 | 14 | 20 | 27 |
| 22. Hwy. 2 mi. W. of Mountain View (SW. ¼ sec. 23, T. 2 (R. 28) | Kimball | Till | Grass | Chernozem | 8 | 10YR2/1 | 5 | 10YR3/3 | ..... | $Cca_1$-11<br>$Cca_2$-14 | 13 |
| 23. 12.5 mi. SE. of Pincher Creek (NW. ¼ sec. 21, T. 4, R. 29) | Kimball | Till | Grass | Chernozem | 12 | 10YR2/2 | 8 | 10YR4/3 | ..... | $Cca_1$-20<br>$Cca_2$-32 | 20 |
| 24. Hwy. 3 mi. W. of Mountain View (NW. ¼ sec. 15, T. 2, R. 28) | Early Continental | Till | Grass | Chernozem | 6 | 10YR2/1 | 5 | 10YR3/3 | 14 | $Cca_1$-6<br>$Cca_2$-12 | 25 |
| 25. Blakiston Canyon, 6 mi. above mouth | Late Wis. Mtn. | Till | Conifers | Podzolic | $A_1$-3<br>$A_2$-1 | 10YR2/2 | .. | ..... | $C_1$-9<br>$C_2$-12 | ..... | 25 |

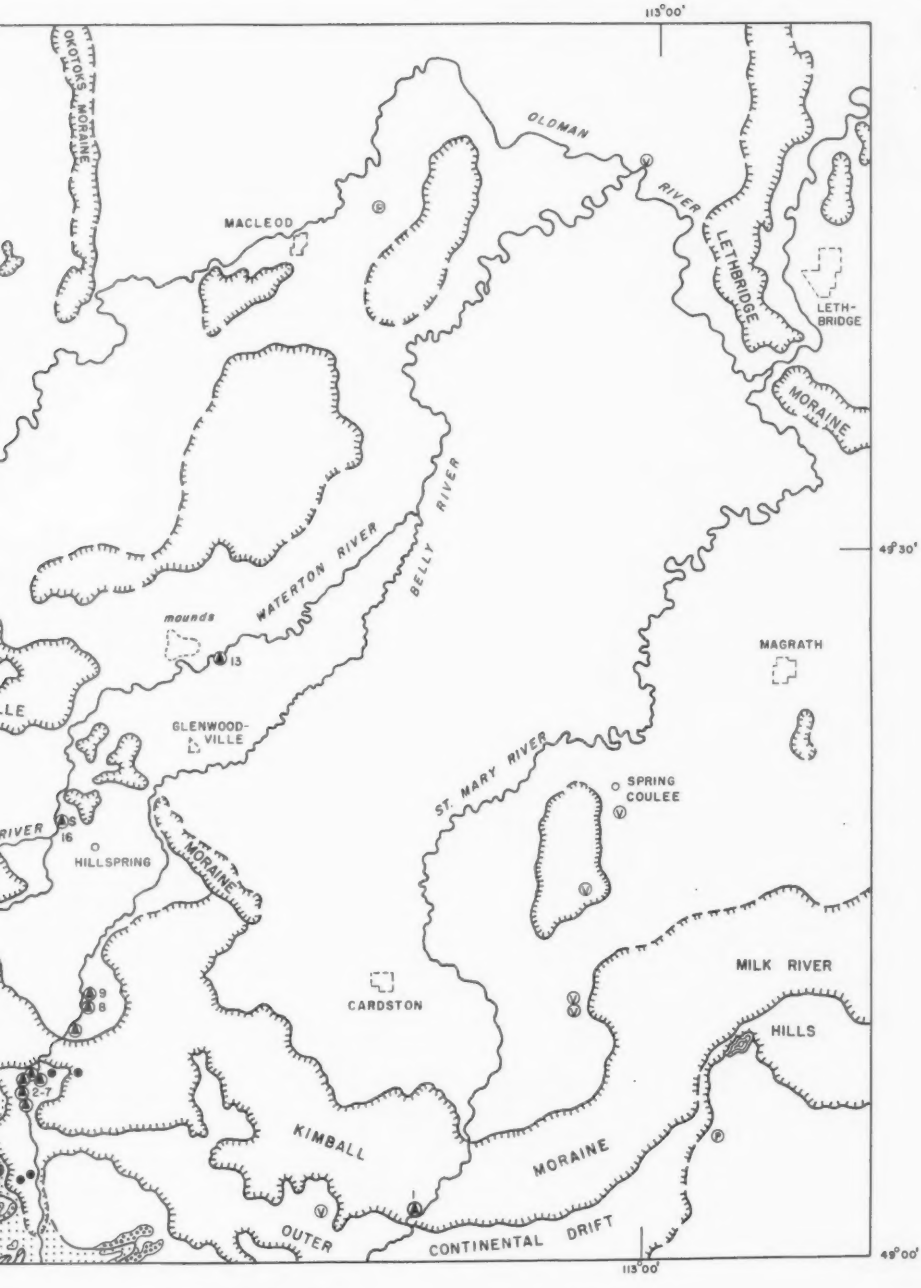
TABLE 6.—SUMMARY OF SOIL-PROFILE OBSERVATIONS (*Continued*)

| Location and locality number (*see* Pl. 8) | Parent material | | Vegetation | Major soil type | *A* Horizon* | | *B* Horizon | | $C_1$ Zone Thickness (Inches) | $C_{ca}$ Zone Thickness (Inches) | Depth of leaching |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | Age | Type | | | Thickness (Inches) | Color† | Thickness (Inches) | Color | | | |
| *Additional profiles—Cont.* | | | | | | | | | | | |
| 26. Gravel pit 3 mi. NE. of Waterton | Late Wis. Mtn. (Waterton) | Gravel | Grass | Chernozem | 6 | 5YR2/2 | 2 | 5YR3/3 | 16 | 22 | 24 |
| 27. Chief Mtn. Hwy. 1 mi. SE. of Waterton River bridge | Late Wis. Mtn. | Till | Grass; aspens | Chernozem | $A_0$-2<br>$A_1$-5 | 10YR3/2<br>10YR2/2 | 2 | 10YR4/3 | 6 | 10 | 23 |
| *Averages:* | | | | | | | | | | | |
| Early Wisconsin Mountain drift | | | | | | | 22.7 | | 29.3 | ..... | 63.7 |
| Late Wisconsin Mountain drift | | | | | | | 2.9 | | 15.9 | 12.5 | 25.2 |
| Continental drift | | | | | | | 4.1 | | 9.6 | 20.0 | 21.6 |

* *See* Plate for zone designations.

† Colors in all cases are of moist specimens determined from Munsell Soil Color Chart.

INDEX MAP TO PLEISTOCENE SECT
FEATURES, VOLCANIC ASH LA




NTAIN DRIFT ○ BURIED PEAT BED Ⓟ PERIGLACIAL STRUCTURES

BURIED SOIL ● SOIL PROFILE LOCATION Ⓥ VOLCANIC ASH LAYER

OCENE SECTIONS, SOIL PROFILES, PERIGLACIAL
NIC ASH LAYER, AND BURIED PEAT BED

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to the Munsell color chart (Table 6). The leached and oxidized $C_1$ (*H*3) horizon ranges in thickness from 20 to 44 inches and averages 29.3 inches.

*Profiles on Late Wisconsin mountain and continental drifts.*—These profiles are so closely alike that no significant age difference is indicated, and they may be treated together as the youngest soil type in the chronosequence. In the mountain drift the *B* horizon has an average thickness of 2.9 inches, the $C_1$ horizon 15.9 inches, the *C*ca horizon 12.5 inches, and the total leached horizon 25.2 inches; whereas, in the continental drift the *B* horizon has an average thickness of 4.1 inches, the $C_1$ horizon 9.6 inches, the *C*ca horizon 20.0 inches, and the total leached zone 21.6 inches. These minor differences are doubtless within the limits of variations due to differences in soil type and parent materials. The continental drift profiles are all Chernozems, whereas, the mountain drift profiles include 3 Prairie, 2 podzolic, and 7 Chernozem profiles. The average depth of leaching, which is 29.2 inches for the Prairie and podzolic soils and 21.9 inches for the Chernozems, suggests that in this regard there may be real differences due to the climatic environment reflected in soil type. However, the slightly thicker *B* horizons and, in part, the shallower leached zones may be explained by the more clayey and less permeable character of the continental drift. Representative described sections of the Late Wisconsin profiles are given in Appendix A.

### *Interpretations*

Significant, major differences in profile development on the various drift sheets of the area are clearly evidenced, even though a limited number of profiles was measured—and there can be no doubt but what the variations are due to differences in age. The ancient paleosol on the Kennedy drift confirms the geomorphic evidence of the antiquity of the drift and points to a Kansan or Nebraskan age. The two younger types of profiles are sharply distinguished and also confirm the geomorphic evidence of age differences. In addition they are closely comparable to Wisconsin soils in the Mississippi Valley and can be used as a basis for correlating the Early Wisconsin mountain drift with the Iowan and Tazewell and the Late Wisconsin mountain and continental drifts with the Cary.

It also appears, in contrast to views commonly held, that some of the pedocal soils, and certainly the Chernozems, show marked differences in textural *B*-horizon development and depths of leaching which can be used as a basis in correlation. A similar view regarding textural *B* horizons in Chernozem, Chestnut, and Brown soils in the Great Plains has been expressed by Thorp and other (1949), who recognize three subgroups in each: (1) *minimal* soils without well-defined *B* horizons, (2) *medial* soils with moderately defined *B* horizons; and (3) maximal soils with strongly defined *B* horizons. On this basis the Early Wisconsin mountain soils are maximal types, and the Late Wisconsin mountain and continental medial, and in a few cases, minimal types.

There is no definite record of major climatic changes in the soil profiles. The Kennedy paleosol, however, is undoubtedly a complex or polygenetic soil, which probably developed largely during the last two interglacials. Only a shallow modern profile is superposed on the old *B* horizon gumbotil. It is of interest that even in this profile there is a zone of lime enrichment indicative of a pedocal soil and relatively dry climatic conditions. The younger Wisconsin soils appear to be dominantly monogenetic, and there is no apparent record of postglacial climatic fluctuations in their profiles.

## Pleistocene Stratigraphy

### *General*

There are numerous sections along the major streams which reveal the superposition of continental drift on mountain drift, as well as other relations, and substantiate the chronology indicated by the areal relations. Most well-exposed sections occur along the St. Mary, Belly, Waterton and Drywood valleys (Pl. 8). The more critical sections are shown graphically in Figures 7, 8, 9, and 10 and are described in the Appendix. A composite section and correlations are shown in Figure 6.

The essential stratigraphic relation of continental till overlying mountain drift in the St. Mary and Belly valleys was noted by both Calhoun (1906, p. 43–49) and Alden (1932, p. 67, 69–72, 115).

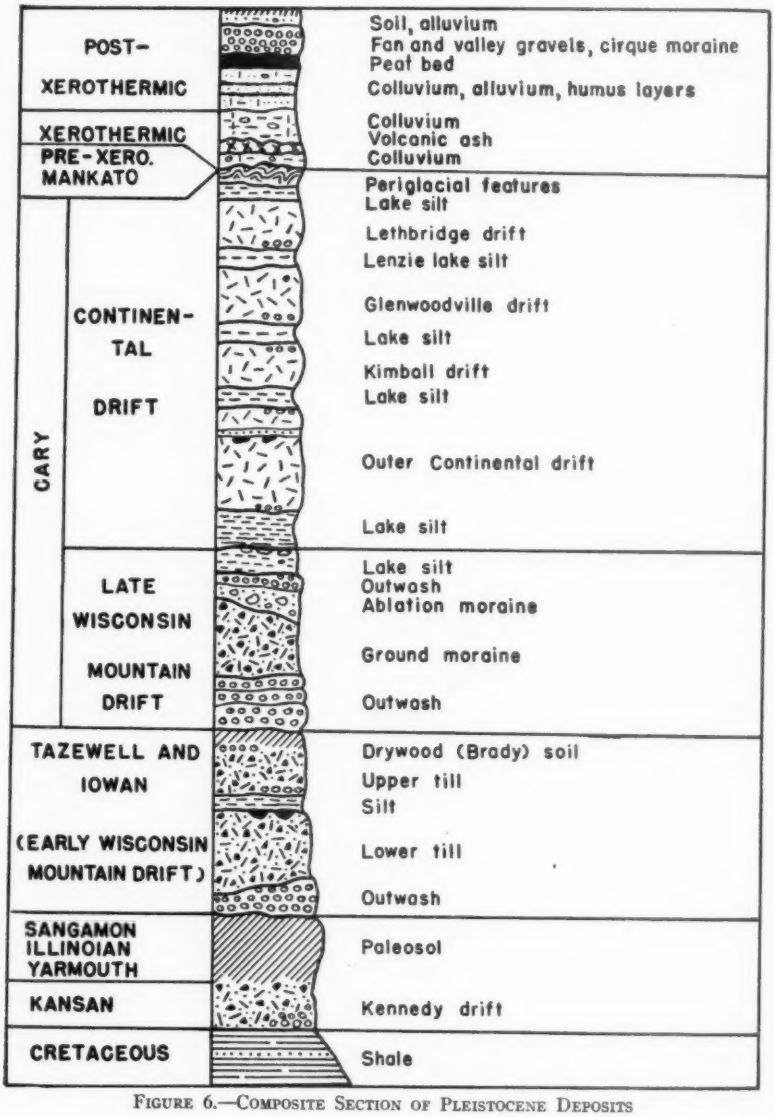



FIGURE 6.—COMPOSITE SECTION OF PLEISTOCENE DEPOSITS

*St. Mary Valley*

The most significant section examined along the St. Mary Valley is the Sloans Ranch section which was first described by Calhoun (1906, p. 48). Later Alden (1932, p. 71–72) discussed the section, although he was unable to find it in the field. The section as described by Calhoun included the following units:

| | Feet |
|---|---|
| "Northeastern till, containing crystalline boulders | 50 |
| Loess, sandy at the top | 6 |
| Soil, of black clayey loam, consisting of the weathered surface of the lower loess bed | 10 |
| Typical loess, containing concretions, hard shells, and bones of small animals | 5 |
| Northeastern drift, containing many decomposed boulders | 2 |
| Shale" | — |

Both Calhoun and Alden were puzzled by the seeming occurrence of weathered pre-Wisconsin drift at a low elevation in the valley.

The section at this locality was fully exposed in 1951, and seemingly Calhoun's section was

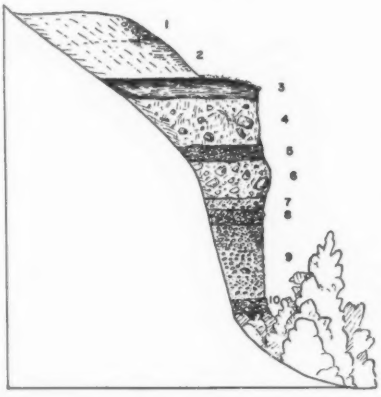


FIGURE 7.—SKETCH OF SLOAN'S RANCH SECTION, ST. MARY VALLEY

(1) Kimball gravel; (2) and (3) glacial Lake St. Mary silts and clays; (4) Outer Continental till; (5) Late Wisconsin mountain outwash; (6) Late Wisconsin mountain till; (7), (8), and (9) Early Wisconsin mountain outwash with possible soil caliche zone at top; (10) till lag from Early Wisconsin mountain till.

limited to the upper part (Fig. 7 and Appendix B, Sect. 1). As reinterpreted the lower "Northeastern" drift is Outer Continental till, the "loesses" and "soil" are lacustrine silts and clays, and the upper "Northeastern" till is Kimball drift. There is no evidence of a weathered zone at the top of the lower till.

Aside from historic interest the section is significant in providing an unusually complete record of Pleistocene events in the area. The mountain drift includes a lower unit of washed till and overlying gravel, which may represent Early Wisconsin drift, and an upper unit of till and gravel, which is referable to the Late Wisconsin. Remnants of an interstadial soil equivalent to the Drywood soil are suggested by an oxidized zone and disseminated lime at the top of the lower unit. The upper part of the section composed of continental drift confirms the chronology based on areal relations and indicates that Glacial Lake St. Mary continued from the time of the retreat from the

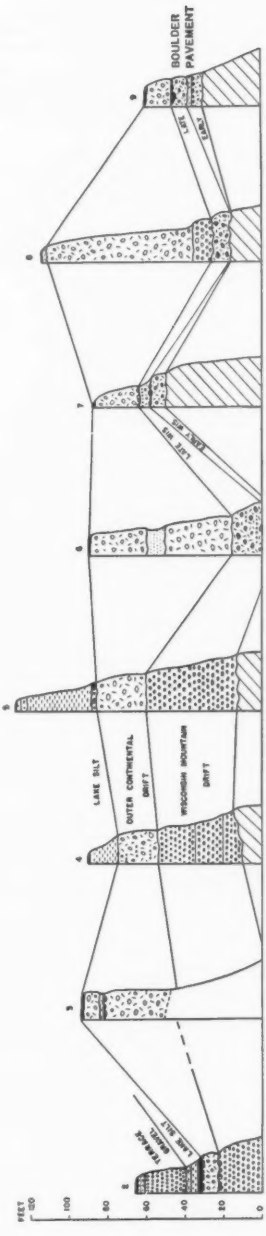



FIGURE 8.—PLEISTOCENE SECTIONS ALONG BELLY RIVER VALLEY (*see* Appendix B for description, Pl. 8 for locations)

Outer Continental stage until the Kimball readvance

*Belly River Valley*

The sections along the Belly River valley extend from localities just outside the Kimball moraine 7 miles downstream (Fig. 8; Pl. 8, secs. 2–9; Appendix B). Taken together they show the following major stratigraphic relations:

(1) The mountain drift again includes two tills which may be correlative with the Early and Late Wisconsin drifts. In section 7 the tills are separated by a prominent zone of lag boulders and in section 9 by a lag deposit of cobbles and boulders overlain by sandy silt. The gravels in sections 2, 4, and 5 cannot be related directly to the tills, but their occurrence in upstream sections indicates that they probably represent retreatal outwash from the ice that deposited the upper till.

(2) The Outer Continental drift in sections 3 and 6 may be separated into two units. This suggests possible complexities which are not apparent from the areal relations.

(3) The lake deposits in sections 2, 4, and 5 record an early stage of Lake Belly River, whereas the outlet gravels in section 2 were deposited after the lake had essentially disappeared.

*Waterton Valley*

The sections in Waterton Valley (Fig. 9) include three sections within the upper valley (nos. 10, 11, 12) and an isolated section north of Glenwoodville about 20 miles downstream (no. 13). The features of major interest are as follows:

(1) The possible two-fold division of the mountain drift is again indicated in section 10 where normal mountain till overlies a partly

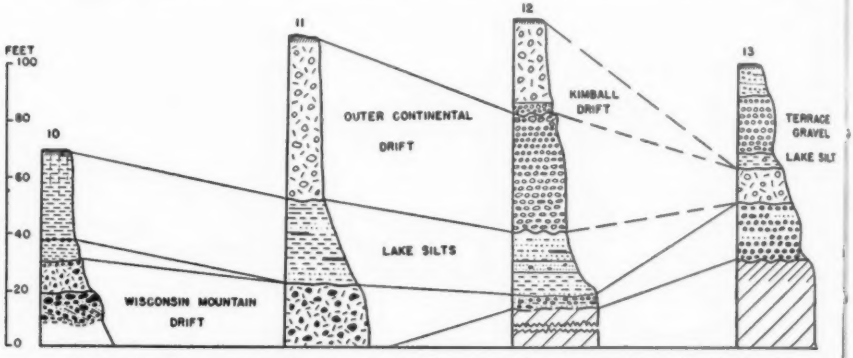



FIGURE 9.—PLEISTOCENE SECTIONS ALONG WATERTON VALLEY (*see* Appendix B, Pl. 8)

indurated boulder lag deposit on coarse gravel. The gravels in section 13 cannot be related directly to the mountain drift upstream and are regarded as Early Wisconsin gravels because of their distance from the mountains and the probable greater extent of Early Wisconsin ice in the piedmont.

(2) The lower lake silts in section 10 are separated from Glacial Lake Waterton silts by a pebble zone indicating a withdrawal of lake waters. This, and the pink color, suggests they were probably deposited in a local basin dammed by Late Wisconsin Waterton moraine rather than in the proglacial lake formed by continental ice.

*Drywood Valley*

This series of sections extends from a point near the center of the Kimball moraine to a locality on Waterton River just below the junction with Drywood River (Fig. 10; Pl. 8; Appendix B, secs. 14, 15, 16). The relations shown by these and other sections are of particular significance because they reveal: (1) a gradation from Early Wisconsin mountain till to outwash at a distance of about 6 miles from the mountain front, and (2) a well-developed buried soil on the top of the mountain drift.

### *"Saskatchewan Gravels"*

The term "Saskatchewan gravels" has been applied widely to gravels underlying the drift in southern Alberta. They were originally named the "South Saskatchewan gravels" from their distribution in the South Saskatchewan drainage basin, but no specific type locality was designated (McConnell, 1885, p. 70). Dawson and McConnell (1895) believed that the gravels represented a pre-Kansan glacial stage, which they named the "Albertan," and were equivalent to early mountain drift near the mountains. Later workers considered them nonglacial deposits reworked from older gravels, and assigned ages varying from late Tertiary to mid-Pleistocene (Calhoun, 1906, p. 49–52; Alden and Stebinger, 1913; Alden, 1932, p. 65; Rutherford, 1937; Horberg, 1952b, p. 309–310).

In the Waterton region the term has been applied to gravels underlying the continental drift along Belly River near Glenwoodville (T. 5, R. 26; Alden and Stebinger, 1913, p. 559) and along St. Mary River near the International Boundary (Alden, 1932, p. 65–67). This raises anew the problem of the gravels, because in both of these areas the regional stratigraphic relations provide strong evidence that they are Early Wisconsin valley-train deposits from mountain glaciers. With increasing distance from the former ice fronts the glacial characteristics disappear, and it is easy to understand how the deposits could be considered nonglacial. This would be especially true of the deposits near Glenwoodville. In this area at least the relations confirm Dawson and McConnell's (1895) original interpretation that the gravels underlying the continental drift are equivalent to mountain drift close to the mountain front. However, it is doubtful that the widespread gravels to the east, which are far from the mountains, represent glacial outwash. Even within the confines of the area studied there are rounded quartzite gravels derived from the mountains—all of "Saskatchewan-type", which occur as terrace gravels on the Flaxville and No. 2 benches in the Milk River Hills, as outwash associated with Kennedy till, and as valley trains of Early and Late Wisconsin age. As with so many deposits of this type, it is increasingly clear that the "Saskatchewan" gravels are diverse in origin and age and that in formal stratigraphic usage the term should be discontinued.

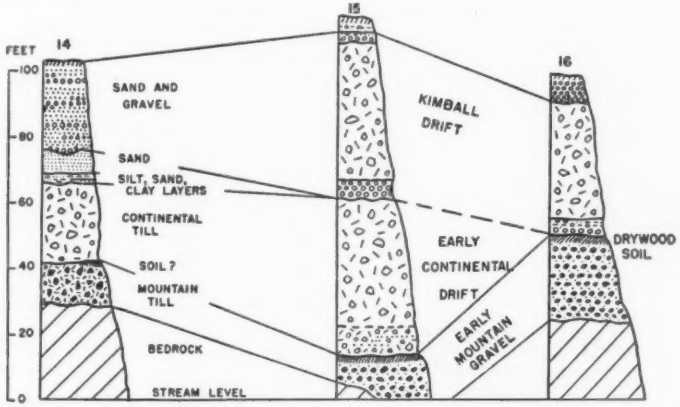



FIGURE 10.—PLEISTOCENE SECTIONS ALONG DRYWOOD CREEK (*see* Appendix B, Pl. 8)

## BURIED SOILS

### *Drywood Soil*

The buried soil on Early Wisconsin mountain gravels below the continental drift along Drywood Creek is here named the Drywood soil, and Section 15 (Appendix B) is designated as the type locality. In addition to the two localities shown in Figure 10, the soil also is well exposed on Drywood Creek 0.7 mile above

locality 15. The horizons in the buried profiles are shown in Figure 11 and are described in detail in Appendix C.

At the type locality the profile is relatively complete and has a modified *B* horizon 20 inches thick and a $C_1$ horizon 6 inches thick, giving a total depth of leaching of 26 inches. The upper 13 inches of less compact loam, designated as $B_1$, may represent a secondary slope-wash deposit composed of mixed *A*-horizon and *B*-horizon materials rather than a primary horizon in the profile (Pl. 9, fig. 1). At locality 16 the *B* horizon is 8 in ches thick and has such a strongly developed clayey texture it can be designated a gumbo-gravel; the $C_1$ horizon is 12 inches thick, and the total depth of leach i ng is 20 inches.

Inasmuch as the profiles have been truncated, it is difficult to compare them with the surface soils. Nevertheless, it appears that the total depth of leaching at least equals that on the Late Wisconsin mountain and contin ental drifts and that the *B* horizon is even more strongly developed. On this basis it seems safe to conclude, in any case, that the time interval recorded by the buried soil would equal or exceed the post-Cary interval believed to be represented by the younger surface soils. This interval, however, does not compare in magnitude with the Sangamon interglacial, which is recorded by thick gumbotils and leached zones in the Mississippi Valley. For this reason the soil is considered interstadial rather than interglacial.

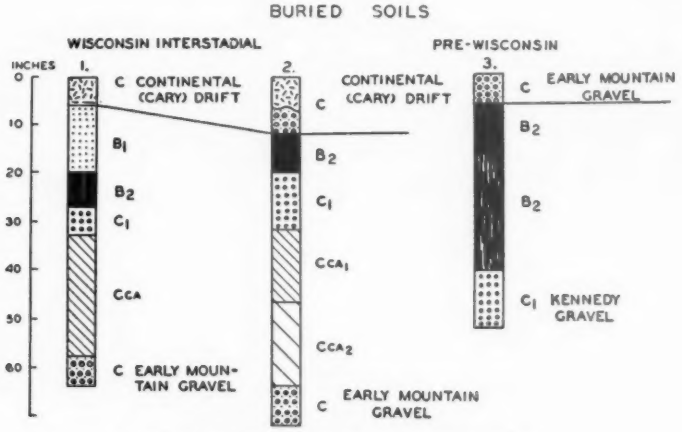



FIGURE 11.—BURIED SOILS (Appendix C; Pl. 8)

### *Buried Soil on Kennedy Drift*

An older buried soil on Kennedy drift below Early Wisconsin mountain gravels is exposed just south of the area on Kennedy Ridge in Montana (Fig. 11; Pls. 3, 9; Appendix C, sec. 3). The section is exposed along the Chief Mountain Highway where it crosses the eastern

PLATE 9.—BURIED SOILS AND BURIED PEAT BED

FIGURE 1.—DRYWOOD SOIL AT TYPE LOCALITY

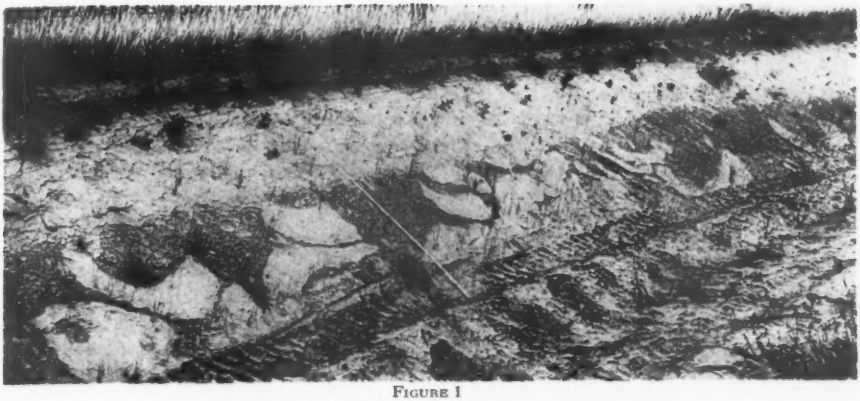
*C*—Outer Continental till; *B*1, *B*2, *C*1, and *Cca* soil horizons on Early Wisconsin mountain gravel, as indicated. Scale shown by 6-inch rule.
cated. Scale shown by 6-inch rule.

FIGURE 2.—PRE-WISCONSIN BURIED SOIL ON KENNEDY RIDGE

*W*—Early Wisconsin mountain gravel; $B2^1$, $B2^2$, and *C*1 horizons on Kennedy mountain drift, as indicated. Scale shown by 6-inch rule.

FIGURE 3.—BURIED PEAT BED ON DUNGARVAN CREEK

*G*—valley gravel; *S*—silt; *P*—peat layer; *T*1—till, leached; *T*2—till, calcareous.



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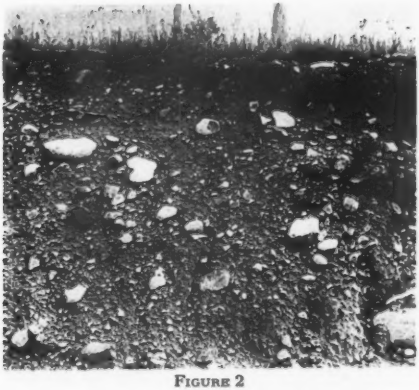

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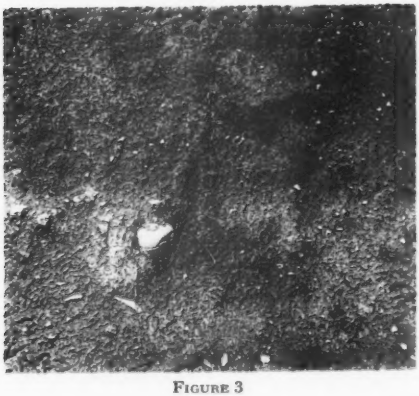

FIGURE 3

BURIED SOIL AND BURIED PEAT BED

end of Kennedy Ridge about 7 miles north-northwest of Babb. The buried profile is incompletely known, as the upper part is truncated and the lower part is not exposed. In the exposed section the *B* horizon is 34 inches thick, and the depth of leaching at least 46 inches. Compared with the more nearly complete surface profile on the Kennedy drift on Two Medicine Ridge, one might conclude that at least 16 feet of *B* horizon had been removed and that the leached zone would continue downward for scores of feet below the base of the exposure. This conclusion, however, is extremely hazardous, because many different histories of weathering probably are recorded by Kennedy soils at various localities. It will suffice to conclude that the buried profile is of greater magnitude than the Drywood soil, and that the regional relations indicate a prolonged interval probably including Yarmouth, Illinoian, and Sangamon time.

## Periglacial Features

### *General*

Periglacial features in the Rocky Mountain piedmont have been noted south of Glacier Park in Montana by Schafer (1949) and in the Lethbridge region, Alberta by Horberg (1952b, p. 315). Those described by Schafer include involutions and fossil ice-wedge structures just outside the limits of Wisconsin continental glaciation, whereas the features in the Lethbridge region are deposits of solifluctional debris in proglacial lake silts of Wisconsin age.

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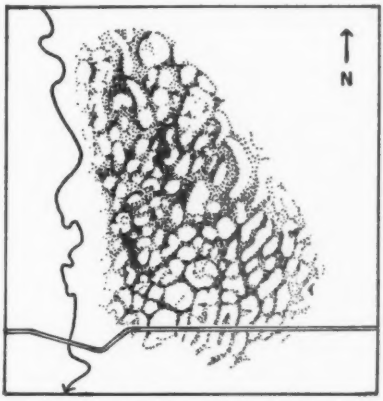
### *Natural Mounds*

The natural mounds occur at the mouth of Scotts Coulee on a deltaic gravel plain related to an early stage of Lake Magrath and cover an area of about 2 square miles.

The individual mounds are 6–80 feet wide with an average of about 40 feet and 1–15 feet high with an average of about 3.5 feet. The intermound depressions are largely undrained

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Most of the common geologic processes can be eliminated from consideration, because of the low slopes of about 3°, the undrained character of the depressions, and the unusual pattern. The

most likely possibilities are: (1) giant interference ripples produced by wave scour; (2) miniature scablands produced by stream floodwaters; and (3) frozen-ground structures (Black, 1952). The arguments pro and con are well known from the long-standing controversy on the origin of the Mima mounds in Washington and will not be reviewed. It appears, however, that serious objections to (1) and (2) are raised by the network pattern of undrained moats and by the absence of associated shore or stream gravels. Therefore, it is concluded that the mound complex is best explained as a system of fossil tundra structures developed under periglacial conditions.

FIGURE 13.—INVOLUTIONS IN FLAXVILLE GRAVELS, MILK RIVER HILLS
Exposed section 10 feet high and 70 feet long. *ST*—silt; *C*—calichified gravel; *GB*—brown, iron-stained gravel; *GD*—deformed gravels of the involuted zone; *GVL*—undisturbed, horizontally bedded gravel (Pl. 8).

### *Involutions near Macleod*

A striking involuted zone in lacustrine silts and clays was exposed in 1952 for a distance of of about 600 feet in a cut along Highway 3, 4 miles east of Macleod (Pl. 8). The involutions occur through a vertical distance of at least 9 feet, the top of the zone being truncated and the base not exposed. Originally the stratigraphic section appears to have included 6 feet or more of dark-gray laminated clay with silt- and fine-sand layers overlain by 3 feet or more of buff, laminated silt with sandy layers and a few clay partings.

The outstanding features of the involuted zone are bulbous masses of silt (brodels), which are without horizontal continuity normal to the face of the exposure and are completely enclosed by clay (Pl. 5, fig. 3). Many silt masses show concentric structures and small-scale breccias composed of splinters of aggregated silt. The stratigraphic relations and the "necks" connecting the brodels to the overlying silts indicate that in the main the silt was intruded downward into the clays rather than the reverse. In places recent desiccation wedges of overlying soil, 2–6 inches in width and up to 2 feet in length, extend downward intersecting the involutions. The modern soil profile clearly truncates the involuted zone.

The involutions are considered periglacial because: (1) they are similar to distinctive periglacial involutions described previously in a voluminous literature (*see* Smith, 1949, for references); (2) they antedate a well-developed soil profile; (3) they extend below the present frost line, which lies at a depth of about 5 feet; and (4) they occur below the surface of a relatively flat lake plain where there has been no opportunity for deformation by mass movement.

### *Involutions in Milk River Hills*

A disturbed and involuted zone in Flaxville gravels is exposed in a gravel pit just outside the continental drift boundary in the southwest part of the Milk River Hills (NE. cor., sec. 36, T. 1, R. 23; Pl. 8). The disturbed zone extends to a depth of more than 10 feet and involves gravels and associated beds of silt and sand. Within the involuted complex there are large brodels of silt and sand, gravel plications, and both small- and large-scale folds (Fig. 13). The gravels are thoroughly weathered, and an old soil caliche is involved in the deformation. However, the modern soil with combined *A* and *B* horizons about 14 inches thick truncates the structures in the disturbed zone. These relations indicate that the deformation was roughly contemporaneous with the Late Wisconsin glaciation.

### *Age of Periglacial Features*

The natural mounds near Glenwoodville and the involuted zone near Macleod postdate glacial lakes Magrath and Macleod respectively

and antedate modern soils which are similar to the post-Late Wisconsin soils elsewhere in the region. With due allowance for the time required for the development of frozen ground after drainage of the lake basins, it is possible the structures were formed at the time of the readvance of the Cary ice to the position of the Lethbridge moraine and during subsequent retreatal stages. Prior to the readvance the ice withdrew to the east at least 50 miles, and undoubtedly a significant time interval was involved (Horberg, 1952b). It seems less likely that the structures were formed during the Mankato stage, because at that time the continental ice at the Altamont moraine was much farther away and glaciers in the mountains were restricted or absent.

The involuted zone in the Milk River Hills, since it lies just outside the continental drift boundary, was probably formed during the local Cary maximum when the ice front was less than 1 mile away.

## Post-Wisconsin Deposits

### *Gravel Fans and Valley Gravels*

Large gravel fans formed by smaller streams debouching from the mountains are the most striking Recent deposits within the region and are represented by: (1) the fans of Sofa Creek and Blakiston Creek in the Lower Waterton Lake basin; (2) Cameron Creek fan on which the town of Waterton is situated; (3) North Fork fan in Belly River valley; (4) Galwey Creek fan; and (5) the fan in upper Yarrow Creek valley (Pl. 1). The fans are grass-covered and partially tree-covered, and the present streams are entrenched in shallow, stabilized channels across them. At present they are subject to slight erosion rather than to aggradation. The small, steep fan 1 mile north of Waterton town site in fact has been completely abandoned, and the stream, entrenched well below the head of the fan, discharges into a small ice-block depression to the north (Pl. 4, fig. 1). Fans of similar type are present in the Glacier Park region in upper Belly River canyon and in St. Mary Valley at the mouths of Kennedy, Swift Current, Wild, and Divide creeks (Alden, 1932, p. 116), and probably at many other localities.

Along the larger streams, equivalent deposits are represented by low gravel terraces, 10–25 feet above stream level, which occur within the canyons in conjunction with minor fans and outside near the mountain front. They are well shown along Blakiston Canyon, upper Cameron Canyon, and lower Yarrow Canyon. In the piedmont area the most impressive example is along Dungarvan Creek where a youthful gravel plain almost 1 mile wide extends eastward 2½ miles from the mountain front. The present stream is confined to a narrow channel about 20 feet below the level of the plain. The gravels in this area overlie the Dungarvan peat bed, described in a later section, which contains a tundra-margin flora and has a radiocarbon age of 3261 ± 250–3327 ± 320 years. On this basis the valley gravels and gravel fans are considered Little Ice Age deposits and equivalent to the cirque moraine in the mountains.

### *Cirque Moraine*

Although there are numerous existing glaciers in Glacier Park, none are known in the Waterton region. However, their presence in recent geologic time is evidenced by an imposing morainic ridge of fresh, unweathered debris at the rim of a large cirque on the northeast side of Mt. Blakiston. The moraine occurs at an elevation of 6000–6500 feet and lies about 1 mile from the back wall of the cirque. It extends completely across the mouth of the cirque and is deeply notched by the present stream. No other examples of moraines of this type were noted in the present study, but it is likely others will be discovered by more detailed investigations.

The moraine is similar to moraines generally assigned to the Little Ice Age (Matthes, 1940) and, as previously noted, is similar in its geomorphic relations to the large gravel fans and low-level valley gravels.

### *Colluvial and Alluvial Deposits*

Colluvium composed mainly of silt with scattered pebbles and rock fragments overlies Pleistocene deposits of all types throughout the area and is present in most exposed sections. It varies in thickness from a few inches to 15 feet but in most places is 2–4 feet thick. The compo-

sition reflects the underlying parent material and slope condition and ranges from clay with scattered pebbles to sandy silt with abundant pebbles and rock fragments. The texture is comcated soil caliche, 26 inches thick, which is marked by disseminated lime and lime-coated pebbles. At this locality the sequence of events was: (1) valley erosion, (2) soil formation on a

TABLE 7.—VOLCANIC ASH

| Locality | Thickness (*inches*) | Depth below Surface (*feet*) | Refractive indices | Particle size |
|---|---|---|---|---|
| (1) Oldman River, 8 miles NW. of Lethbridge (center, NW. ¼ sec. 26, T. 9, R. 23) | 4 | 6 | 1.507–.002 | silt |
| (2) South Pincher Creek, 0.5 mile S. of town of Pincher Creek (SE. corner, NE. ¼ sec. 15, T. 6, R. 30) | 3 | 3 | 1.515–.005 | fine sand |
| (3) Alberta Railway and Irrigation Canal, 1.25 miles S. of Spring Coulee (NW. ¼ sec. 21, T. 4, R. 23) | 6 | 8 | ..... | ..... |
| (4) Same, 6.5 miles S. of Spring Coulee (Center, N. ½ sec. 31, T. 3, R. 23) | 8 | 8 | 1.505–.005 | silt |
| (5) Same, 3 miles E. of Woolford Station (SE. corner sec. 1, T. 3, R. 23) | 6 | 11 | 1.518–.005 | silt |
| (6) Boundary Creek, 2 miles N. of Carway (NW. corner sec. 11, T. 1, R. 26) | 3 | 12 | 1.507–.005 | silt |
| (7) Galata, Montana, 0.75 mile W. on Hwy. 2 | 15 | 9 | 1.510–.005 | silt |

monly loesslike, and in certain sections thin loess may be present. Stratification is generally absent or rudely developed. In many sections, particularly in the upper part, two or more buried humus layers are present, and in a few sections bone fragments and gastropod shells were noted.

Although colluvial deposition is going on at present in restricted localities, essentially all of the extensive colluvium was deposited earlier and under different environmental conditions. This is evidenced by: (1) a youthful soil profile with leached zones of 12–20 inches developed on the deposits; (2) the occurrence of deposits in stream banks and valley slopes that are now being actively eroded; and (3) the widespread occurrence on upland slopes that are now completely stabilized.

In many places significant valley entrenchment and at least in one place soil development preceded colluvial deposition. In lower Dungarvan Valley (NW. ¼, NW. ¼, sec. 23, T. 3, R. 29), for example, colluvium extends down the valley slope with increasing thickness to a graded surface about 75 feet below the upland and 30 feet above the present stream. On the till surface below the colluvium there is a truncstabilized till slope, (3) erosion of the soil zone on the slope, (4) colluvium deposition, (5) recent soil development on the upper slopes and old valley floor, and (6) stream entrenchment. Except for the buried soil, these relations are consistent throughout the area and suggest a definite correlation with postglacial climatic fluctuations. In addition, the occurrence of humus layers in the upper part of many sections suggests a two-fold division and a change in climatic conditions during colluvium deposition.

In valleys, colluvium and alluvium commonly occur together, and there is little doubt but what there is lateral gradation from one to the other. The alluvium commonly consists of stratified sandy silt and silty sand and in places contains considerable organic material. A volcanic ash layer of wide regional extent occurs in both types of deposits.

### *Volcanic Ash*

A volcanic ash layer in postglacial deposits along the Oldman River 9 miles northwest of Lethbridge was noted in a previous investigation (Horberg, 1952a; 1952b, p. 317). In the present study the ash was recognized at five

additional localities within the region (Pl. 8) and at one locality in Montana. The specific localities, depth of occurrences, and properties of the glass are given in Table 7. Similarities in occurrences, refractive indices, shape of shards, and plagioclase composition (andesine-labradorite) clearly indicate that the ash layers are correlative. At most localities the ash occurs in the lower part of the postglacial deposits, and at localities 4 and 5 (Table 7) it underlies the upper part of the colluvium, which contains humus layers (Appendix D, 1).

Postglacial volcanic ash zones in bog deposits have been observed by Hansen (1948) in southwestern Glacier National Park and from localities farther north in the general vicinity of Edmonton, Alberta (Hansen, 1949a; 1949b). They are dated from pollen analyses as belonging to the postglacial Xerothermic period and tentatively attributed to eruptions from Glacier Peak in north-central Washington. Glass fragments from the Glacier Park locality, insofar as determinable from clouded shards, have indices of refraction of about 1.515, which is similar to indice values obtained for the glass in the Waterton region. This, and the similarity in geologic occurrence, suggests that a single, widespread ash fall may be represented.

### *Buried Peat Bed*

A buried peat bed overlain by valley gravels of a 20-foot terrace is exposed along Dungarvan Creek at a single locality 2½ miles from the mountain front (Pl. 8; Pl. 9, fig. 3). The measured section is described in Appendix D, 2. As previously noted, the valley gravels, which form a prominent valley plain traceable to the canyon mouth, are considered correlative with the large gravel fans found elsewhere in the piedmont.

The peat bed is 2 feet thick and is composed largely of rudely stratified or massive grayish-brown to blackish-brown peat which grades into laminated, calcareous silt above and below. Flattened wood fragments up to 2 inches in diameter are common in the peat and also as scattered fragments in the associated silts. In places the sequence is deformed by slump movements along low-angle faults. Although the peat layer is exposed continuously for 100 feet, it pinches out at the east end of the exposure. These relations, as well as the character of the deposits, indicate deposition in a local pond and bog. However, deposition was preceded by soil development and erosion so that the deposits record only a small part of postglacial time.

Samples of the peat and wood fragments were studied by L. R. Wilson of the University of Massachusetts with the following results (personal communication, 1951):

"1. The wood is spruce and appears to be *Picea mariana* (black spruce) or small specimens of *Picea glanca* (white spruce)....
2. The peat contains an abundance of pollen making up the following spectrum:

| | |
|---|---|
| *Picea Mariana* (black spruce) | 5% |
| *Picea glauca* (white spruce) | 12 |
| *Pinus Banksiana* (jack pine) | 74 |
| *Abies* sp. (fir) | 2.5 |
| Grass sp. | 6.5 |

In addition I also found: 1 *Sphagnum* moss spore; 1 composite pollen (*Cichorieae*), contamination; 1 *Trichothyrites* sp., conifer parasite; 1 *Rhizophagites* sp., fungus spore, conifer parasite; several other types of fungus spores; many leaf fragments of aquatic Hypnaceous mosses; many Diflugia tests (Protozoans), northern species; many aquatic invertebrate cysts.

I have checked these results with some of the forestry maps and find the trees mentioned as pollen fossils to be living today quite a distance north of the peat locality. The most significant thing about the pollen spectrum is the total absence of deciduous tree pollen. The ecology suggested by the spectrum is now existent at the edge of the tundra in northern Canada."

Radiocarbon age determinations by Professor Libby of the University of Chicago (Libby, 1952, p. 96, samples 606, 607) give an age of 3327 ± 320 years for a sample of peat and an age of 3261 ± 250 years for a wood specimen. These determinations place the materials in the early part of the post-Xerothermic period (Little Ice Age) of postglacial time and suggest that the valley gravels overlying the peat may represent the climax of cooler conditions and a local glacial climate.

### *Chronology*

The post-Wisconsin deposits of the region show a consistent development, which, together with relations to the volcanic ash layer and buried peat bed, suggest the following tentative postglacial chronology:

end of Kennedy Ridge about 7 miles north-northwest of Babb. The buried profile is incompletely known, as the upper part is truncated and the lower part is not exposed. In the exposed section the *B* horizon is 34 inches thick, and the depth of leaching at least 46 inches. Compared with the more nearly complete surface profile on the Kennedy drift on Two Medicine Ridge, one might conclude that at least 16 feet of *B* horizon had been removed and that the leached zone would continue downward for scores of feet below the base of the exposure. This conclusion, however, is extremely hazardous, because many different histories of weathering probably are recorded by Kennedy soils at various localities. It will suffice to conclude that the buried profile is of greater magnitude than the Drywood soil, and that the regional relations indicate a prolonged interval probably including Yarmouth, Illinoian, and Sangamon time.

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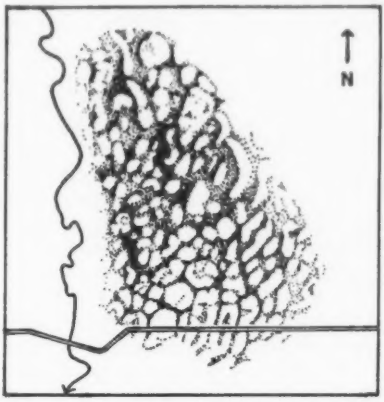


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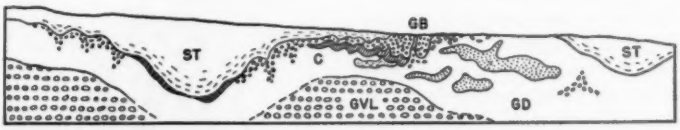


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Along the larger streams, equivalent deposits are represented by low gravel terraces, 10–25 feet above stream level, which occur within the canyons in conjunction with minor fans and outside near the mountain front. They are well shown along Blakiston Canyon, upper Cameron Canyon, and lower Yarrow Canyon. In the piedmont area the most impressive example is along Dungarvan Creek where a youthful gravel plain almost 1 mile wide extends eastward 2½ miles from the mountain front. The present stream is confined to a narrow channel about 20 feet below the level of the plain. The gravels in this area overlie the Dungarvan peat bed, described in a later section, which contains a tundra-margin flora and has a radiocarbon age of 3261 ± 250–3327 ± 320 years. On this basis the valley gravels and gravel fans are considered Little Ice Age deposits and equivalent to the cirque moraine in the mountains.

### *Cirque Moraine*

Although there are numerous existing glaciers in Glacier Park, none are known in the Waterton region. However, their presence in recent geologic time is evidenced by an imposing morainic ridge of fresh, unweathered debris at the rim of a large cirque on the northeast side of Mt. Blakiston. The moraine occurs at an elevation of 6000–6500 feet and lies about 1 mile from the back wall of the cirque. It extends completely across the mouth of the cirque and is deeply notched by the present stream. No other examples of moraines of this type were noted in the present study, but it is likely others will be discovered by more detailed investigations.

The moraine is similar to moraines generally assigned to the Little Ice Age (Matthes, 1940) and, as previously noted, is similar in its geomorphic relations to the large gravel fans and low-level valley gravels.

### *Colluvial and Alluvial Deposits*

Colluvium composed mainly of silt with scattered pebbles and rock fragments overlies Pleistocene deposits of all types throughout the area and is present in most exposed sections. It varies in thickness from a few inches to 15 feet but in most places is 2–4 feet thick. The compo-

sition reflects the underlying parent material and slope condition and ranges from clay with scattered pebbles to sandy silt with abundant pebbles and rock fragments. The texture is commonly loesslike, and in certain sections thin loess may be present. Stratification is generally absent or rudely developed. In many sections, particularly in the upper part, two or more buried humus layers are present, and in a few sections bone fragments and gastropod shells were noted.

Table 7.—Volcanic Ash

| Locality | Thickness (*inches*) | Depth below Surface (*feet*) | Refractive indices | Particle size |
|---|---|---|---|---|
| (1) Oldman River, 8 miles NW. of Lethbridge (center, NW. ¼ sec. 26, T. 9, R. 23) | 4 | 6 | 1.507–.002 | silt |
| (2) South Pincher Creek, 0.5 mile S. of town of Pincher Creek (SE. corner, NE. ¼ sec. 15, T. 6, R. 30) | 3 | 3 | 1.515–.005 | fine sand |
| (3) Alberta Railway and Irrigation Canal, 1.25 miles S. of Spring Coulee (NW. ¼ sec. 21, T. 4, R. 23) | 6 | 8 | ..... | ..... |
| (4) Same, 6.5 miles S. of Spring Coulee (Center, N. ½ sec. 31, T. 3, R. 23) | 8 | 8 | 1.505–.005 | silt |
| (5) Same, 3 miles E. of Woolford Station (SE. corner sec. 1, T. 3, R. 23) | 6 | 11 | 1.518–.005 | silt |
| (6) Boundary Creek, 2 miles N. of Carway (NW. corner sec. 11, T. 1, R. 26) | 3 | 12 | 1.507–.005 | silt |
| (7) Galata, Montana, 0.75 mile W. on Hwy. 2 | 15 | 9 | 1.510–.005 | silt |

Although colluvial deposition is going on at present in restricted localities, essentially all of the extensive colluvium was deposited earlier and under different environmental conditions. This is evidenced by: (1) a youthful soil profile with leached zones of 12–20 inches developed on the deposits; (2) the occurrence of deposits in stream banks and valley slopes that are now being actively eroded; and (3) the widespread occurrence on upland slopes that are now completely stabilized.

In many places significant valley entrenchment and at least in one place soil development preceded colluvial deposition. In lower Dungarvan Valley (NW. ¼, NW. ¼, sec. 23, T. 3, R. 29), for example, colluvium extends down the valley slope with increasing thickness to a graded surface about 75 feet below the upland and 30 feet above the present stream. On the till surface below the colluvium there is a truncated soil caliche, 26 inches thick, which is marked by disseminated lime and lime-coated pebbles. At this locality the sequence of events was: (1) valley erosion, (2) soil formation on a stabilized till slope, (3) erosion of the soil zone on the slope, (4) colluvium deposition, (5) recent soil development on the upper slopes and old valley floor, and (6) stream entrenchment. Except for the buried soil, these relations are consistent throughout the area and suggest a definite correlation with postglacial climatic fluctuations. In addition, the occurrence of humus layers in the upper part of many sections suggests a two-fold division and a change in climatic conditions during colluvium deposition.

In valleys, colluvium and alluvium commonly occur together, and there is little doubt but what there is lateral gradation from one to the other. The alluvium commonly consists of stratified sandy silt and silty sand and in places contains considerable organic material. A volcanic ash layer of wide regional extent occurs in both types of deposits.

### *Volcanic Ash*

A volcanic ash layer in postglacial deposits along the Oldman River 9 miles northwest of Lethbridge was noted in a previous investigation (Horberg, 1952a; 1952b, p. 317). In the present study the ash was recognized at five

additional localities within the region (Pl. 8) and at one locality in Montana. The specific localities, depth of occurrences, and properties of the glass are given in Table 7. Similarities in occurrences, refractive indices, shape of shards, and plagioclase composition (andesine-labradorite) clearly indicate that the ash layers are correlative. At most localities the ash occurs in the lower part of the postglacial deposits, and at localities 4 and 5 (Table 7) it underlies the upper part of the colluvium, which contains humus layers (Appendix D, 1).

Postglacial volcanic ash zones in bog deposits have been observed by Hansen (1948) in southwestern Glacier National Park and from localities farther north in the general vicinity of Edmonton, Alberta (Hansen, 1949a; 1949b). They are dated from pollen analyses as belonging to the postglacial Xerothermic period and tentatively attributed to eruptions from Glacier Peak in north-central Washington. Glass fragments from the Glacier Park locality, insofar as determinable from clouded shards, have indices of refraction of about 1.515, which is similar to indice values obtained for the glass in the Waterton region. This, and the similarity in geologic occurrence, suggests that a single, widespread ash fall may be represented.

### *Buried Peat Bed*

A buried peat bed overlain by valley gravels of a 20-foot terrace is exposed along Dungarvan Creek at a single locality 2½ miles from the mountain front (Pl. 8; Pl. 9, fig. 3). The measured section is described in Appendix D, 2. As previously noted, the valley gravels, which form a prominent valley plain traceable to the canyon mouth, are considered correlative with the large gravel fans found elsewhere in the piedmont.

The peat bed is 2 feet thick and is composed largely of rudely stratified or massive grayish-brown to blackish-brown peat which grades into laminated, calcareous silt above and below. Flattened wood fragments up to 2 inches in diameter are common in the peat and also as scattered fragments in the associated silts. In places the sequence is deformed by slump movements along low-angle faults. Although the peat layer is exposed continuously for 100 feet, it pinches out at the east end of the exposure. These relations, as well as the character of the deposits, indicate deposition in a local pond and bog. However, deposition was preceded by soil development and erosion so that the deposits record only a small part of postglacial time.

Samples of the peat and wood fragments were studied by L. R. Wilson of the University of Massachusetts with the following results (personal communication, 1951):

"1. The wood is spruce and appears to be *Picea mariana* (black spruce) or small specimens of *Picea glanca* (white spruce). . . .
2. The peat contains an abundance of pollen making up the following spectrum:

| | |
|---|---|
| *Picea Mariana* (black spruce) | 5% |
| *Picea glauca* (white spruce) | 12 |
| *Pinus Banksiana* (jack pine) | 74 |
| *Abies* sp. (fir) | 2.5 |
| Grass sp. | 6.5 |

In addition I also found: 1 *Sphagnum* moss spore; 1 composite pollen (*Cichorieae*), contamination; 1 *Trichothyrites* sp., conifer parasite; 1 *Rhizophagites* sp., fungus spore, conifer parasite; several other types of fungus spores; many leaf fragments of aquatic Hypnaceous mosses; many Difugia tests (Protozoans), northern species; many aquatic invertebrate cysts.

I have checked these results with some of the forestry maps and find the trees mentioned as pollen fossils to be living today quite a distance north of the peat locality. The most significant thing about the pollen spectrum is the total absence of deciduous tree pollen. The ecology suggested by the spectrum is now existent at the edge of the tundra in northern Canada."

Radiocarbon age determinations by Professor Libby of the University of Chicago (Libby, 1952, p. 96, samples 606, 607) give an age of 3327 ± 320 years for a sample of peat and an age of 3261 ± 250 years for a wood specimen. These determinations place the materials in the early part of the post-Xerothermic period (Little Ice Age) of postglacial time and suggest that the valley gravels overlying the peat may represent the climax of cooler conditions and a local glacial climate.

### *Chronology*

The post-Wisconsin deposits of the region show a consistent development, which, together with relations to the volcanic ash layer and buried peat bed, suggest the following tentative postglacial chronology:

| | Years ago |
|---|---|
| Post-Xerothermic (Little Ice Age).... | 0–4000 |
| Recent erosion and minor alluviation | |
| Climax-deposition of cirque moraines, gravel fans, and valley gravels | |
| Deposition of buried peat bed (3300 B.P.); deposition of colluvium and alluvium with humus layers | |
| Xerothermic..... | 4000–7000 |
| Maximum colluvium deposition; volcanic ash fall; erosion | |
| Pre-Xerothermic..... | 7000–8000 |
| Soil development | |
| Mankato..... | 8000–11000 |
| Valley erosion; intensified frost action | |

## Correlation of Rocky Mountain Pleistocene Deposits

The mountain glacial deposits in the Waterton region conform closely with the geomorphic pattern established at an early date by Blackwelder (1915, p. 307–340) for mountain drifts in the Middle Rockies of western Wyoming. On this basis the Late Wisconsin moraines are correlated with the Pinedale of Wyoming, the Early Wisconsin moraines with the Bull Lake, and the Kennedy drift remnants with the Buffalo.

The regional relations, soils, and geomorphic expression of the continental drift in the Waterton region in addition provide strong evidence for the Cary age of both the Late Wisconsin mountain moraines and the overlapping continental drift. This conclusion is based on the following lines of evidence:

(1) The regional relations place the continental drift of the area far in front of the Altamont moraine, which is generally considered to mark the outer Mankato boundary (Bretz, 1941, p. 34; Flint *et al.*, 1945), and behind outer continental drift in Montana tentatively assigned to the Iowan or Iowan and Tazewell (Flint *et al.* 1945). Thus both the continental drift and the Late Wisconsin mountain drift, which is overlapped by continental drift, are considered Cary.

(2) The soils, which are similar on both drifts, have an average depth of leaching of 2.0 feet compared with: 2.5 feet for the post-Mankato of Iowa (Kay, 1931); 3.5 feet for the post-Cary of Indiana (Thornbury, 1940); 4.5 feet for the post-Tazewell of Indiana (Thornbury, 1940), and 5.5 feet for the post-Iowan of Iowa (Kay, 1931). Hence, a Cary rather than a Tazewell or Iowan age is indicated.

(3) The geomorphic expression of the continental and Late Wisconsin mountain moraines is similar to that of the Cary and Mankato rather than the Tazewell of the Mississippi Valley region. Initial ridges, knobs, and depressions are largely unmodified by postglacial processes, drainage integration is slight, and hundreds of small lakes and ponds are present. This again, in light of (1), points to a Cary age.

Inasmuch as the mountain moraines are similar to the Pinedale and other moraines in the Rockies that have generally been considered Mankato (*see* Moss, 1951, for a recent summary), the Cary dating of these moraines is one of the most significant general conclusions arising from the present investigation. As a corollary it would appear that in the northern Rockies at least the Mankato glaciers were greatly restricted or absent.

The correlation of the Early Wisconsin mountain moraines with the Iowan and Tazewell is in line with earlier correlations of the Bull Lake and other similar moraines and conforms to the twofold division of the Wisconsin now generally accepted in the continental interior. By this two-fold division an early Wisconsin phase consisting of Tazewell, Iowan, and possibly Farmdale (Leighton and Willman, 1950, p. 602) is separated by the Brady soil from a late Wisconsin phase consisting of Cary and Mankato (Ruhe, 1952; Frye and Leonard, 1952, p. 132-135). Substantial evidence, based on differential development of soil profiles and the buried Drywood soil occupying the same general stratigraphic position as the Brady, supports the geomorphic evidence for this correlation.

In the Glacier Park-Waterton region both the extent of postglacial erosion and the profound weathering of the Kennedy drift strongly support correlation with the Kansan rather than the Illinoian. In fact it is impossible to duplicate the advanced stage of either in older drifts of the Mississippi Valley region. Canyon cutting to depths of 1500 feet, and reduction of the drift sheet to isolated patches occurred since deposition of the Kennedy. In the Mississippi Valley region both the Kansan and Illinoian occur as continuous, extensive drift sheets. The Illinoian, however, still retains recognizable

morainic forms and initial till plains which place it closer in time to the Wisconsin than to the Kansan. In spite of important differences in the erosional histories of the two regions, the geomorphic evidence indicates a Kansan rather than an Illinoian age. This conclusion also is supported by the ancient paleosol on the Kennedy drift of Two Medicine Ridge, Glacier Park. In this profile the *B* horizon (gumbotil) is 19 feet thick compared with mean thicknesses on post-Illinoian soils of 4.5 feet in Indiana (Thornbury, 1940), and the leached zone is 61 feet thick compared with mean thicknesses in post-Illinoian soils of 13.2 feet in Indiana (Thornbury, 1940). The Kennedy paleosol compares more favorably with buried Yarmouth and Aftonian soils in Iowa: the Yarmouth has a *B* horizon (gumbotil) with a mean thickness of 11.0 feet and a mean depth of leaching of about 30 feet; the Aftonian has a *B* horizon with a mean thickness of 8.5 feet and a mean depth of leaching of about 20 feet (Kay, 1931). Although numerous uncertainties are involved in comparisons of this type, the Kennedy paleosol clearly represents more than post-Illinoian time.

The absence of deposits logically assignable to the Illinoian stage also conforms, with few exceptions, to previous studies elsewhere in the Rocky Mountains. The general reluctance to recognize the Illinoian may be due in part to the absence of Illinoian drift from the Keewatin center in the Mississippi Valley, but more largely it is due to the profound geomorphic unconformity between drifts commonly assigned to the Wisconsin stages and a pre-Wisconsin stage. Although changes in the course of the Missouri River in central South Dakota have been attributed to Illinoian glaciation (Warren, 1952; Crandall, 1953), as yet there is no published stratigraphic evidence to support the recognition of Illinoian drift of Keewatin origin. If Illinoian drift were ever deposited, it was probably so restricted in its extent that it was obliterated by Sangamon valley cutting and Wisconsin glaciation.

The correlation of the Kennedy drift with the Kansan rather than Nebraskan or Kansan and Nebraskan remains uncertain. Certainly the Two Medicine Ridge paleosol does not prohibit a Nebraskan age. However, drift remnants of Kennedy type in the Rocky Mountains are generally assigned to the Kansan rather than the Nebraskan, and to date it is only in the Sierra Nevada that two ancient drifts (McGee and Sherwin) are recognized and tentatively correlated with the Kansan and Nebraskan (Blackwelder, 1931, p. 91B). It should be recalled also that Alden (1932, p. 31–32), in the Glacier Park region, considered the possibility that the pre-Wisconsin (Kennedy) drift on Number 1 and 2 benches respectively represented Nebraskan and Kansan. Correlation with the Kansan rather than Nebraskan is favored because of the greater likelihood of preservation of Kansan drift and the record of major mid-Pleistocene erosion which would have destroyed any Nebraskan drift present or reduced it to isolated patches not readily differentiated from the Kennedy.

In general, the evidence available suggests synchroneity of glacial maxima in the mountains with Kansan, Iowan-Tazewell, and Cary maxima of the Keewatin ice sheet in the western part of the continental interior. The Mankato could be considered an exception, although it should be recalled that the Mankato boundary on the west (Altamont moraine) was greatly restricted as compared with the Cary.

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Department of Geology, University of Chicago 37, Illinois
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## Appendix A.—Representative Soil Profiles

*see* Plate 7 for graphic sections and horizon designations

*No. 3—Profile on Late Wisconsin mountain till, Belly River lobe; Crooked Creek and Chief Mountain Highway; podzolic; grass*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| $A_1$–*H*1 | Loam, silty, dark grayish brown (10 YR 3/2), loose | 5 | 5 |
| $B$–*H*2 | Loam, dark yellowish brown (10 YR 4/4), denser than above | 4 | 9 |
| $C_1$–*H*3 | Till, light brown to purplish gray, leached and oxidized, less dense than above | 17 | 26 |
| *C*–*H*4 | Till as above, but calcareous | | |

*No. 11—Profile on Early Wisconsin mountain drift, Sofa Creek, west lateral moraine; podzolic; conifers*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| $A_0$–*H*1 | Mat of forest litter, dark brown (10 YR 2/2) | 2 | 2 |
| $A_2$–*H*1 | Silt, powdery, grayish brown (10 YR 4/2), rootlets | 8 | 10 |
| $B$–*H*2 | Loam, clayey, strong brown (10 YR 5/6), angular, blocky, dense, hard; insoluble pebbles with iron-oxide stain; reddish brown, iron-stained zone in lower 4 inches | 28 | 38 |
| $C_1$–*H*3 | Till, stony, brown to purplish gray, leached and oxidized | 44 | 82 |
| *C*–*H*4 | Till, as above, weakly calcareous; strongly calcareous till below | 30 | 112 |

No. 15—*Profile on Kimball till, highway cut North of Cloudy Ridge (SW. ¼ sec. 26, T.3, R.30); podzolic; grass*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| $A_1$–*H*1 | Loam, silty, dark grayish brown (10 YR 3/2), loose | 5 | 5 |
| $B$–*H*2 | Loam, clayey, dark yellowish brown (10 YR 4/4), firm, rude columnar structure | 3 | 8 |
| $C_1$–*H*3 | Till, clayey, yellowish brown, leached | 21 | 29 |
| *C*–*H*4 | Same, calcareous | ? | ? |

No. 24—*Profile on Outer Continental till, highway cut along Cardston road east of Belly River (NW. ¼ sec. 15, T.2, R.28); Chernozem; grass*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| $A_1$–*H*1 | Loam, black (10 YR 2/1), loose, rootlets | 6 | 6 |
| $B$–*H*2 | Loam, dark brown (10 YR 3/3), firm | 5 | 11 |
| $C_1$–*H*3 | Till, yellowish brown, more friable than above, leached | 14 | 25 |
| $Ca_1$–*H*4$ca_1$ | Till, brownish gray, pebbly; lime veinlets and coatings on pebbles disappear toward top | 18 | 43 |
| $Ca_2$–*H*4$ca_2$ | Till, as above with discontinuous veinlets and coatings | 48 | 91 |
| *C*–*H*4 | Till, pebbly, yellowish brown, calcareous | ? | ? |

*No. 25—Profile on Late Wisconsin mountain till, canyon moraine in Blakiston Canyon, 6 miles above mouth at Red Rock Canyon camp; conifers; podzolic*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| $A_0$–*H*1 | Loam, silty, very dark brown (10 YR 2/2), rootlet mat | 3 | 3 |
| $A_2$–*H*1 | Silt, light gray, powdery | 1 | 4 |
| $C_1$–*H*3 | Till, pebbly, reddish brown (5 YR 4/3), leached, oxidized | 9 | 13 |
| | Same, yellow brown to purplish gray, leached, partly oxidized | 12 | 25 |
| *C*–*H*5 | Till, stoney, purplish gray, calcareous | | |

No. 27—*Profile on Late Wisconsin mountain till, Waterton moraine (SE. ¼ sec. 9, T.2, R.29); Chernozem; grass*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| $A_0$–*H*1 | Loam, dark grayish brown (10 YR 3/2), few pebbles, loose | 2 | 2 |
| $A_2$–*H*1 | Loam, very dark brown (10 YR 2/2), few pebbles, loose | 3 | 5 |
| $B$–*H*2 | Loam, dark brown (10 YR 4/3), few pebbles, firm | 2 | 7 |
| $C_1$–*H*3 | Till, pebbly, light brown, leached | 6 | 13 |
| *Ca*–*H*4*ca* | Till, pebbly, light brownish gray, lime dissemination and coatings on pebbles | 10 | 23 |
| *C*–*H*4 | Till, pebbly, brownish gray | ? | ? |

No. 28—*Profile on Kennedy drift, Two Medicine Ridge at old lookout station just east of Highway 2 (SW. ¼ sec. 16, T.32 N., R. 13 W.), Glacier Park, Montana*

| Horizon | Description | Thickness (feet) | Depth (feet) |
|---|---|---|---|
| $A_1$–*H*1 | Grass and cedar mat, pebble lag | 0.5 | 0.5 |
| $A_2$–*H*1 | Clay, pebbly, reddish gray (5 YR 5/2), noncalcareous | 1.5 | 2 |
| $B_1$–*H*2 | Clay, pebbly, reddish brown (5 YR 4/4), plastic, noncalcareous, more pebbly than above | 3 | 5 |
| $B_2$–*H*2 | Clay, pebbly, yellow red (5 YR 5/8), very plastic, noncalcareous; quartzite and argillite insolubles; no igneous or carbonate rocks; could be termed a "gumbo cobble gravel" | 14 | 19 |
| $C_1$–*H*3 | Loam, stony, clayey, yellow red (5 YR 4/6) slightly plastic, noncalcareous except for relicts of carbonate boulders in lower part; diorites from Belt series decomposed | 42 | 61 |
| *Cca*–*H*4ca | Tillite with conglomerate layers, purplish gray, calcareous cement; contains, in order of abundance: argillite, quartzite, limestone, dolomite, sandstone, and diorite; striae common on argillites | 98 | 159 |
| *C*–*H*5 | Till, purplish gray, partly indurated, calcareous; many striated stones; varved clay inclusions | 20 | 179 |
| *C*–*H*5 | Varved clay and silt, purplish gray, rafted pebbles | 10 | 189 |

## Appendix B.—Pleistocene Drift Sections

*see* Plate 8 for locations

No. 1—*Section opposite Sloan's Ranch on north side of St. Mary River (SE. corner SW. ¼ sec. 16, T.1, R.25; Fig. 7)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Kimball outwash: | | |
| Gravel, pebbles and cobbles, sand matrix, poorly exposed | 15 | 15 |
| Glacial Lake St. Mary deposits: | | |
| Clays and silts, buff to gray, slumped, poorly exposed | 10 | 25 |
| Clay gray, laminated, contorted, calcareous, rare pebbles | 2 | 27 |
| Silt, buff, loesslike texture, obscure contorted laminae, limey concretions, calcareous, clay and sand partings, pebbles abundant near base | 3 | 30 |
| Outer Continental till: | | |
| Till, dark gray, calcareous, clayey, crystallines | 8 | 38 |
| Late Wisconsin mountain drift: | | |
| Gravel, cobbles in sandy silt matrix, buff, calcareous | 3 | 41 |
| Till, bouldery, large angular blocks in gray, clayey matrix with red argillite chips, gravelly in places; largely Belt rocks with some local Cretaceous, no crystallines | 7 | 48 |
| Early Wisconsin (?) mountain drift: | | |
| Sand, silty, with cobbles and pebbles, buff, oxidized, disseminated caliche, partly indurated | 2.5 | 50.5 |
| Gravel, poorly sorted, tillike, buff, oxidized, disseminated caliche | 3 | 53.5 |
| Gravel, pebbly, gray, unoxidized; imbrication and foresets indicate eastward flow | 3.5 | 57 |
| Gravel, bouldery lag, sand matrix | 5 | 62 |
| Cretaceous sandstone and shale | | |

No. 2—*Belly River, east side, 2 miles W. of Payne Lake (SW. ¼ sec. 9, T.2, R.28; Fig. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Postglacial colluvium: | | |
| Colluvium, silty, pebbly, calcareous, buried humus zones | 5 | 5 |
| Terrace gravel: | | |
| Gravel, silty, poorly sorted, average 1 inch | 2 | 7 |
| Glacial Lake Belly River deposits: | | |
| Sand and silt, interstratified, buff, calcareous; varves with silt and clay layers half an inch to 2 inches, buff to pink, calcareous | 4 | 13 |
| Clay, finely laminated, dark gray | 2 | 15 |
| Outer Continental drift: | | |
| Gravel and bouldery lag | 1 | 16 |
| Till, buff, clayey, crystallines | 6 | 22 |
| Same, dark gray | 2 | 24 |
| Late Wisconsin mountain drift: | | |
| Gravel, purplish sand matrix, average about 3 inches, maximum 10 inches, Belt lithology | 20 | 44 |

No. 3—*Belly River, west side (NW. ¼ sec. 8, T.2, R.28; Fig. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Outer Continental drift: | | |
| Till, dark brownish gray, calcareous, friable | 8 | 8 |
| Sand, silty, grayish brown, cobble lag at base | 1 | 9 |
| Till, dark gray, calcareous, very hard and dense | 30 | 39 |

No. 4—*Belly River, east side, 0.7 miles SW. of Cardston highway bridge (NE. ¼ SW. ¼ sec. 17, T.2, R.28; Fig. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Glacial Lake Belly River silt: | | |
| Silt, buff, laminated, poorly exposed | 15 | 15 |
| Outer Continental till: | | |
| Till, clayey, grayish brown, gravel layers | 20 | 35 |
| Late Wisconsin mountain gravel: | | |
| Gravel, coarse, bouldery, gray | 20 | 55 |
| Same, purplish gray | 15 | 70 |
| Same, brownish gray | 10 | 80 |
| Cretaceous shale | | |

No. 5—*Belly River, south side, at Cardston highway bridge (NE. ¼ sec. 17, T.2, R.28; Fig. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Glacial Lake Belly River silts: | | |
| Silt, finely stratified, sand and clay layers, buff, calcareous | 33 | 33 |
| Outer Continental drift: | | |
| Gravel and bouldery lag, discontinuous | 3 | 36 |
| Till, clayey, cobbly, buff, calcareous, crystallines | 26 | 62 |
| Late Wisconsin mountain drift: | | |
| Gravel, bouldery, poorly sorted, sandy matrix, purplish gray, calcareous, Belt lithology | 47 | 109 |
| Cretaceous shale | | |

No. 6—*Belly River, south side, 0.3 mile below Cardston highway bridge (SE. ¼ sec. 20, T.2, R.28; Fig. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Outer Continental drift: | | |
| Till, gravelly, grayish brown, abundant crystallines | 25 | 25 |
| Sand, gray, well sorted | 10 | 35 |
| Till, clayey, dark gray | 35 | 70 |
| Late Wisconsin Mountain till: | | |
| Till, stony, purplish gray, gravel layers; to river level | 15 | 85 |

No. 7—*Belly River, east side, 0.5 miles east of Cardston highway bridge (SE. ¼ SW. ¼ sec. 21, T.2, R.28; Fig. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Outer Continental till: | | |
| Till, clayey, yellowish brown, crystallines | 10 | 10 |
| Late Wisconsin mountain drift: | | |
| Boulder lag | 1 | 11 |
| Gravel, silty, red clay laminae | 1 | 12 |
| Till, stony, yellowish brown, Belt lithology | 4 | 16 |
| Early Wisconsin (?) mountain drift: | | |
| Boulder lag | 1 | 17 |
| Till, clayey, purplish brown | 8 | 25 |
| Cretaceous shale | | |

No. 8—*Belly River, east side (center, SE. ¼ sec. 1, T.3, R.28; Fig. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Glacial Lake Caldwell deposit: | | |
| Silt, buff, calcareous, ¼-inch laminae | 3 | 3 |
| Outer Continental drift: | | |
| Till, clayey, buff to gray; pebbly at base | 75 | 78 |
| Gravel, sandy, average about 2 inches, crystallines | 10 | 88 |
| Early Wisconsin mountain drift: | | |
| Till, purplish gray, Belt lithology | 9 | 97 |
| Cretaceous bedrock | | |

No. 9—*Belly River, gully on east side (SW. ¼, NE. ¼, sec. 1, T.3, R.28; Fig. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Outer Continental till: | | |
| Till, clayey, brownish gray, crystallines | 14 | 14 |
| Late Wisconsin mountain drift: | | |
| Boulder pavement; facetted boulders 2–3 feet in diameter with uniformly oriented striae indicate continental ice moved S. 12° W. | 2 | 16 |
| Till, pebbly, silty, purplish gray | 9 | 25 |
| Silt, sandy, pebbly, buff, calcareous | 1 | 26 |
| Early Wisconsin (?) mountain drift: | | |
| Boulder and cobble lag, silty sand matrix | 2 | 28 |
| Till, silty, pebbly, purplish buff, calcareous, much denser than till above | 3 | 31 |
| Cretaceous sandstone | | |

No. 10—*Crooked Creek, 1 mile NE. of Waterton River bridge (N.E. ¼ SW. ¼ sec. 16, T.2, R.29; Fig. 9)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Glacial Lake Waterton silt: | | |
| Soil profile on silt | 3.5 | 3.5 |
| Silt, some sand laminae, buff, poorly stratified | 30 | 33.5 |
| Local lake silt: | | |
| Silt, sand, and clay laminae, pink, laminated; parting of angular gravels at top | 7 | 40.5 |
| Late Wisconsin mountain drift: | | |
| Gravel, discontinuous | 0.5 | 41 |
| Till, bouldery, gravelly, pinkish gray | 11 | 52 |
| Early Wisconsin (?) mountain drift: | | |
| Gravel and bouldery lag, reddish-gray sand and silt matrix, partly indurated, largely horizontal bedding but some cross beds; imbrication indicates eastward flow | 9 | 61 |

No. 11—*Waterton River, west side (SE. ¼ SW. ¼ sec. 1, T.3, R.29; Fig. 9)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Outer Continental till: | | |
| Till, clayey, brownish gray, crystallines | 58 | 58 |
| Glacial Lake Waterton deposit: | | |
| Silt and clay, laminated, pink to gray | 28 | 86 |
| Late Wisconsin mountain till: | | |
| Till, gravelly, purplish gray, Belt lithology | 22 | 108 |

No. 12—*Waterton River, east side; at point where Kimball moraine front crosses river (SW. corner sec. 7, T.3, R.28; Fig. 9)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Kimball drift: | | |
| Till, clayey, grayish buff, abundant crystallines | 30 | 30 |
| Gravel, gray, becomes coarser from bottom to top | 4 | 34 |
| Outer Continental drift: | | |
| Boulder and cobble lag | 2 | 36 |
| Gravel, sandy, gray, average about 2 inches | 40 | 76 |
| Glacial Lake Waterton deposit: | | |
| Sand, fine, silty, laminated; some pink clay laminae | 10 | 86 |
| Sand, grayish brown, isolated pebbles and cobbles | 4 | 90 |
| Silt, sandy, finely laminated, some clay laminae | 8 | 98 |
| Gravel, silty, poorly exposed | 4 | 102 |
| Cretaceous sandstone | | |

No. 13—*Waterton River, south side, 5 miles N. of Glenwoodville, (NE., ¼ sec. 31 T.6, R.26; Fig. 9)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Postglacial colluvium: | | |
| Silt, pebbly, with buried humus layers | 6 | 6 |
| Terrace deposit: | | |
| Gravel, coarse horizontally bedded, maximum 10 inches | 15 | 21 |
| Glacial Lake silt: | | |
| Silt, laminated, buff, discontinuous | 5 | 26 |
| Continental till: | | |
| Till, clayey, yellowish brown, some dark-gray unoxidized till, crystallines, sand and gravel beds | 12 | 38 |
| Early Wisconsin mountain drift: | | |
| Gravel, purplish gray, sand matrix, average about 3 inches, maximum 14 inches, horizontally bedded | 20 | 58 |
| Cretaceous shale: | | |
| Shale, yellowish brown, oxidized | 8 | 66 |
| Shale, dark gray | 30 | 96 |

No. 14—*Drywood Creek, south side, 0.8 mile SW. of Pincher Creek highway bridge (SE. ¼ SW. ¼ sec. 18, T.4, R.29; Fig. 10*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Postglacial colluvium and soil | 3 | 3 |
| Kimball drift: | | |
| Sand and gravel, interbedded, grayish brown | 26 | 39 |
| Sand, medium to coarse, buff, laminated | 7 | 46 |
| Glacial lake deposit: | | |
| Silt and fine sand, clay laminae | 3 | 49 |
| Outer Continental till: | | |
| Till, clayey, dark gray, crystallines | 25 | 74 |
| Early Wisconsin mountain drift: | | |
| Till, gravelly, purplish gray; yellowish-brown to reddish-brown oxidized till exposed in places may represent Drywood soil; poorly exposed | 12 | 86 |
| Cretaceous shale | | |

No. 15—*Drywood Creek, north side, near junction with Narrow Creek (SE. ¼ NE. ¼ sec. 15. T.4, R.28; Fig. 10)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Postglacial colluvium: | | |
| Silt, pebbly, dark grayish brown | 5 | 5 |
| Kimball drift: | | |
| Gravel, coarse, sandy, grayish brown | 3 | 8 |
| Till, yellowish brown, poorly exposed | 41 | 49 |
| Gravel, coarse, grayish brown | 6 | 55 |
| Outer Continental till: | | |
| Till, clayey, dense, dark gray | 38 | 93 |
| Till, gravelly, dark gray, sand layers | 8 | 101 |
| Drywood paleosol on Early Wisconsin mountain gravels | 4 | 105 |
| Early Wisconsin mountain drift: | | |
| Gravel, purplish gray, sand matrix, rude horizontal bedding, partly indurated | 12 | 117 |

No. 16—*Waterton River, east side, 0.5 mile N. of Hill Spring bridge (SE. ¼ sec. 26, T.4, R.28; Fig. 10)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Terrace deposit: | | |
| Gravel, sandy, poorly exposed | 8 | 8 |
| Outer Continental drift: | | |
| Till, clayey, dark gray, calcareous | 35 | 43 |
| Gravel, silty, yellowish brown, calcareous | 0.5 | 43.5 |
| Drywood soil on Early Wisconsin mountain drift: | | |
| Soil profile | 4.5 | 48 |
| Early Wisconsin mountain drift: | | |
| Gravel, purplish-gray sand matrix, calcareous | 20 | 68 |
| Cretaceous sandstone and shale | | |

## Appendix C.—Buried Soil Profiles

No. 1—*Drywood soil, type locality.—Drywood Creek (Profile 1, Fig. 11; sec. 15, Appendix B; SE. ¼ NE. ¼ sec. 15, T.4, R.28)*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| | Outer Continental till: | | |
| | Till, clayey, dark gray, calcareous; 1-inch, purplish-gray clay layer at base | | |
| | Drywood soil on Early Wisconsin mountain outwash: | | |
| $B_1$ | Loam, clayey, stony, dark grayish brown, angular, blocky; noncalcareous except for some veinlets of secondary lime; insoluble residuals of quartzite and rare argillite; possibly includes redeposited *A*- and *B*-zone materials | 13 | 13 |
| $B_2$ | Loam, clayey, stony, yellowish brown, vertical foliation, scattered insoluble particles, weakly calcareous in places | 7 | 20 |
| $C_1$ | Loam, less clayey and dense, yellowish brown, friable; some pebbles with remnants of basal lime coating; weakly calcareous in places | 6 | 26 |
| Cca | Gravel, grayish brown, calcareous; lime coatings on many pebbles; disseminated lime; limestone and dolomite pebbles | 25 | 51 |
| *C* | Gravel, purplish gray, sand matrix, boulders up to 14 inches, rude horizontal bedding | | |

No. 2—*Drywood soil.—Waterton River, 0.5 mile N. of Hill Springs bridge (Profile 2, Fig. 11; sec. 16, Appendix B; SE. ¼ sec. 26, T.4, R.28)*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| Continental drift: | | | |
| | Till, clayey, dark gray, calcareous, 5-inch layer of silty gravel at base | ? | ? |
| Drywood soil on Early Wisconsin mountain gravel: | | | |
| $B_2$ | Gumbogravel, clayey, yellowish brown, insoluble residuals, noncalcareous except for secondary lime | 8 | 8 |
| $C_1$ | Loam, gravelly, less clayey and more pebbly than above, yellowish brown, noncalcareous except for secondary lime | 12 | 20 |
| $Cca_1$ | Gravel, grayish brown, disseminated lime and lime-coated pebbles, limestone and dolomite pebbles | 15 | 35 |
| $Cca_2$ | Gravel, grayish brown, discontinuous disseminated lime, some lime-coated pebbles, transitional above and below | 18 | 53 |
| $C$ | Gravel, purplish-gray sand matrix, calcareous | | |

No. 3—*Pre-Wisconsin buried soil.—Kennedy Ridge, cut along Chief Mountain Highway about 7 miles N.-NW. of Babb, Montana (Fig. 11)*

| Horizon | Description | Thickness (inches) | Depth (inches) |
|---|---|---|---|
| Early Wisconsin mountain drift: | | | |
| | Gravel, purplish-gray sand matrix, average about 4 inches, maximum 2 feet, 9 feet thick | ? | ? |
| Pre-Wisconsin soil on Kennedy drift: | | | |
| $B_{21}$ | Silt, clayey, scattered, pebbles, yellowish brown (10 YR 5/6), insoluble residual pebbles, noncalcareous except for some secondary lime | 10 | 10 |
| $B_{22}$ | Same, not so clayey and dense, more pebbly | 24 | 34 |
| $C_1$ | Gravel, reddish brown (5 YR 4/4), iron-stained, many soft weathered pebbles, noncalcareous except for some secondary lime, average about 2 inches, base not exposed | 12 | 46 |

## Appendix D.—Sections of Postglacial Deposits

No. 1—*Volcanic ash section.—Alberta Railway and Irrigation Canal, 3 miles east of Woolford Station (SE. corner sec. 1, T.3, R.24)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Postglacial deposits: | | |
| Humus, brown | 0.25 | 0.25 |
| Silt, pebbly, sandy, dark gray, rudely stratified; alluvium | 1.5 | 1.75 |
| Humus, black | 0.4 | 2.15 |
| Silt, clayey, sandy layers, light grayish brown, rudely stratified; alluvium | 2.0 | 4.15 |
| Humus, black | 0.5 | 4.65 |
| Silt, as above | 3.5 | 8.15 |
| Humus, black | 0.5 | 8.65 |
| Clay, dark gray, scattered pebbles, unstratified, blocky; colluvium | 3.5 | 12.15 |
| Volcanic ash, purplish gray, discontinuous | 0.5 | 12.65 |
| Clay, as above | 1.5 | 14.15 |
| Glacial lake deposit: | | |
| Silt and clay, varved | 3.0 | 17.15 |
| Kimball till: | | |
| Till, clayey, yellowish brown | 6.0 | 23.15 |

No. 2—*Dungarvan peat bed.—Dungarvan Creek (NW. ¼ NW. ¼ sec. 8, T.2, R.29, Pl. 8)*

| Deposits | Thickness (feet) | Depth (feet) |
|---|---|---|
| Postglacial deposits: | | |
| Soil zone | 1 | 1 |
| Gravel, gray, disseminated lime, iron-stained in places | 12 | 13 |
| Silt, laminated, gray, sand and clay laminae, wood fragments, gastropods | 4 | 17 |
| Peat, dark brown to black, laminated in places, wood fragments, scattered pebbles, grades into silt above and below | 2 | 19 |
| Silt, pebbly, gray, laminated, calcareous, humus streaks | 2 | 21 |
| Truncated soil profile: | | |
| Till, brown, noncalcareous; in places there is a 3-inch, lime-enriched zone at base | 1 | 22 |
| Outer Continental till: | | |
| Till, clayey, brownish gray, calcareous | 8 | 30 |

*ppendix*

| Depth (inches) |
|---|
| ? |
| 8 |
| 20 |
| 35 |
| 53 |

*N.-NW.*

| Depth (inches) |
|---|
| ? |
| 10 |
| 34 |
| 46 |

*on (SE.*

| Depth (feet) |
|---|
| 0.25 |
| 1.75 |
| 2.15 |
| 4.15 |
| 4.65 |
| 8.15 |
| 8.65 |
| 12.15 |
| 12.65 |
| 14.15 |
| 17.15 |
| 23.15 |

| Depth (feet) |
|---|
| 1 |
| 13 |
| 17 |
| 19 |
| 21 |
| 22 |
| 30 |